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[American Journal of Science, Vol. 257, Summer 1959, P. 465-480]

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SUMMER 1959

DEPTH OF LEACHING IN RELATION TO CARBONATE CONTENT OF TILL IN CENTRAL NEW YORK STATE

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ABSTRACT. Depth of leaching varies inversely with calcium carbonate content of till in a north-south transect in the uplands of the Finger Lakes region of central New York. Variability of slope and of till texture and density were limited by careful selection of sampling sites to minimize factors which might mask the relationship of carbonate content to depth of leaching.

The transect is marked by progressive southward decrease in carbonate content and increase in depth of leaching. The most striking change in carbonate content, depth of leaching and soil profile characteristics occurs in a narrow transition zone which lacks topographic features indicative of a significant glacial stillstand. Instead, this pattern of change is apparently related to distance from outcrops of calcareous rocks and to currents of major ice movement.

Differences in age of the drift within the range of the transect are not great enough to have an effect on the depth of leaching comparable to the effect of carbonate content. For a distance on both sides of the Valley Heads drift border the upland till is characterized by deep leaching and low carbonate content.

INTRODUCTION

Geologists in north-central and northeastern United States have long recognized a correlation between assumed age of glacial drift and depth of carbonate removal by leaching. Young drift sheets contain carbonate at depths nearer the surface than do similarly exposed drift sheets which are older and therefore have probably been subjected to longer weathering and soil development.

Pedologists, on the other hand, consider differences in depth of leaching among soils developed from Wisconsin drift to be related to differences in initial carbonate content rather than to age. They have used depth of leaching as a criterion of presumed lithological differences between parent materials of the soil.

This paper reports on a quantitative study of the relationship between leaching and initial carbonate content along a north-south transect in upland drift of the Finger Lakes region of New York. Along the transect, depth of leaching ranges between wide extremes on parent materials of similar texture, density and exposure but dissimilar carbonate content. Carbonate content decreases and depth of leaching increases consistently southward away from limestone exposures. Whereas deep leaching and low carbonate content characterize upland drift for a distance north and south of the Valley Heads moraine, the sharpest changes in depth of leaching, carbonate content and soil-profile characteristics occur in a narrow transition zone that lacks topographic evidence of a glacial stillstand.

This investigation underscores the need for careful appraisal of effects of

initial carbonate content in all cases where correlation of Wisconsin drifts is based only on comparative depths of leaching.

LITERATURE REVIEW

Difference in depth of carbonate removal by ground water was first used by Alden and Leighton (1917) as a basis for distinguishing between Iowan and Kansan drifts of northeastern Iowa. Since that time many other investigators in the midwest have employed this criterion to distinguish drifts of the several Pleistocene stages (Kay, 1931; Kay and Graham, 1943; Leighton, 1923; Leighton and MacClintock, 1930; Schultz, Leuninghoener and Frankforter, 1951; Thornbury, 1940). Under favorable situations the method is usable even for separating substages of the Wisconsin.

A number of geologists have discussed the several factors that may influence depth of leaching of drift (Bryan and Albritton, 1943; Denny, 1956; Dreimanis, 1957; Flint, 1949; Howard, 1946; Leighton and MacClintock, 1930; MacClintock, 1954a, 1954b; Willman and Payne, 1942). Soil scientists, too, concerned with carbonate removal as one aspect of soil development, have stressed factors other than age which may control rate of leaching (Bradfield, 1941; Chandler, 1941; Ellis, 1938; Jenny and Leonard, 1934; Russell and Engle, 1925). Climate, vegetation, topographic position, depth to water table, carbonate content, ratio of magnesium to calcium carbonate, permeability and stratification of parent material—all these have been shown to affect the rate of carbonate removal.

In New York, MacClintock (1954a, 1954b) attempted to subordinate complicating factors to age by sampling only in permeable gravel on flat benchlands well above valley bottoms. He found that depths of carbonate removal grouped themselves in three distinct ranges which he characterized as showing "knee-high", "head-high" and "more than head-high" leaching, respectively. Sites with 10 percent to 35 percent carbonate pebbles occur in each of the three groups of leaching depths, showing that the clustering of the data is at least partly independent of variation in initial carbonate content. On this basis, MacClintock inferred that three Wisconsin drifts are represented in New York. Denny (1956) and Dreimanis (1957) recently suggested that depth of leaching should be analyzed in relation to total free carbonate, rather than to carbonate content of the pebble fraction alone, as in MacClintock's data.

Soil scientists in central and eastern North America describe significant relationships between soil-profile characteristics and carbonate content of parent material (Allen and Whiteside, 1954; Conrey, et al., 1937; Ehrlich and Rice, 1955; Ehrlich, Rice and Ellis, 1955). The work of Cline and McCaleb (Cline, 1953, 1955; McCaleb, 1954; McCaleb and Cline, 1950) in New York indicates that the distribution of certain soils is related to difference in carbonate content of the parent material. They further suggest that this geographic arrangement may be paralleled by a chronologic succession of profiles as carbonate is removed by leaching.

Pedologic studies cited above show that duration of exposure to weathering may be less important than initial carbonate content in controlling depth of leaching, but published data afford little basis for quantitative evaluation of this hypothesis. The present study was undertaken to combine geologic and

pedologic concepts and techniques in analyzing the role of initial carbonate content in controlling the depth of leaching.

GEOLOGICAL SETTING

East and south of Cayuga Lake in the Finger Lake region of central New York (fig. 1) the Allegheny Plateau is underlain by gently dipping sedimentary rocks, with progressively older strata exposed northward toward the Ontario lake plain. Glacial drift partly fills deeply scoured valleys and blankets the uplands.

The composition of the glacial drift reflects the lithology of bedrock over which the Pleistocene ice sheets moved (Holmes, 1952). Each lithologic constituent of the drift is at maximum abundance immediately south of the source exposure, and becomes progressively more scarce southward through attrition and dilution. The Lower Devonian Onondaga limestone which crops out at the north edge of the study area, is a major source of carbonate in the upland drift sampled in this transect. Stratigraphically overlying the Onondaga limestone are middle Devonian strata consisting primarily of shale with a number of thin limestone members, including the Tully limestone. These calcareous sediments are well exposed in the walls of Cayuga trough and crop out west of the south end of Owasco Lake (fig. 1). South of this line the area is underlain by a monotonous succession of shale, siltstone and sandstone units with uniformly low carbonate content.

Drift of pre-Wisconsin age is preserved only in drift-filled interglacial gorges (von Engeln, 1929, 1931; Holmes, 1935; Muller, 1957a). The surface drift can be conveniently divided into older and younger Wisconsin drift (Denny, 1956) separated by the Valley Heads moraine (Fairchild, 1932).

South of the Valley Heads moraine the drift on the upland is thin and barely masks the bedrock. In extensive areas the weathering profile penetrates downward through the drift into underlying bedrock. Calcareous till is difficult to locate. The drift contains almost no fragments of crystalline or carbonate rock, but is composed of material of local derivation. Constructional topography is limited in extent and modified in form.

In the lowlands south of the Valley Heads moraine, ice contact stratified drift contrasts in several respects with the adjacent upland drift. The valley fill is thick and consists primarily of stratified drift with little till. The gravel is well sorted and clean in appearance and includes numerous far-traveled cobbles of crystalline and of carbonate rock. Because of the coarse texture of this drift, constructional topographic features are prominent and little modified.

MacClintock and Apfel (1944) correlated the bright, clean drift south of the moraine in the Susquehanna River drainage with similar deposits along the drift border in western New York. They called these deposits Binghamton drift, distinguishing it from the Olean drift which emerges from beneath Binghamton deposits west of Salamanca, New York. They recognized, but did not emphasize the fact that drift resembling the Olean occurs on uplands both north and south of east-west trending valleys filled with Binghamton drift in south central New York. Tarr (1909), failing to distinguish end moraines in the lower Chemung and Chenango Valleys, concluded that the bright gravels are stagnant-ice deposits laid down during protracted recession from the Wis-

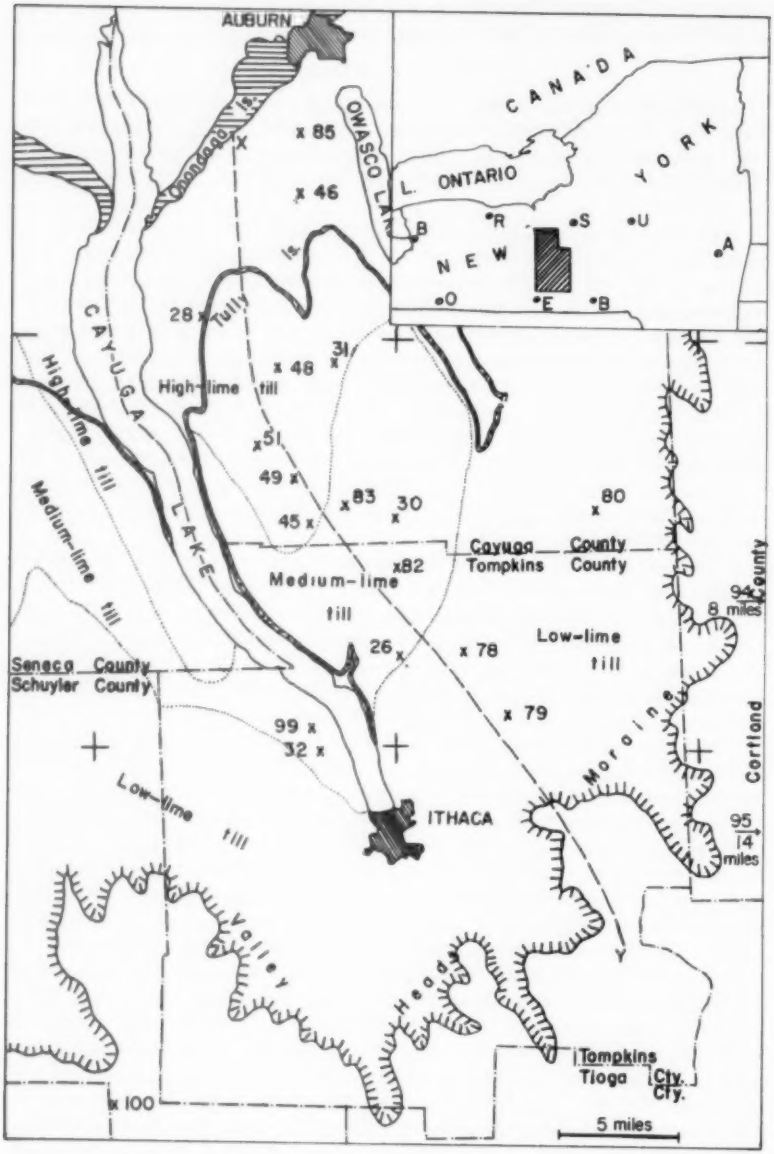

Fig. 1. Index map showing county lines, Tully and Onondaga limestone belts, Valley Heads moraine, sampling sites, transect projection line X-Y, and low-, medium- and high-lime till zones.

consin drift border. Denny (1956) concurred in this interpretation and suggested that the Binghamton border may pass under the Valley Heads moraine in the Finger Lakes region.

In view of the unresolved status of the Binghamton deposits south of the Finger Lake region, the present writers prefer to retain the term Binghamton in a descriptive sense only, without necessary inference of correlation or time equivalence to drift west of the Salamanca re-entrant.

The last Wisconsin glaciation to reach the area of this study left distinct marginal features and massive moraines, the most notable among which have been designated the Valley Heads moraine (Fairchild, 1932) (fig. 1). This has been interpreted as an end moraine deposited following minor readvance of the ice sheet (Tarr, 1909; von Engeln, 1938). In many of the through valleys of central New York, the moraine forms the divide between St. Lawrence and Susquehanna drainage systems. In such positions the moraine is a massive accumulation of coarse, poorly sorted gravel and of sandy or silty till (von Engeln, 1921, 1938). It exhibits rough kame-and-kettle topography with steep-sided hills and small depressions which may contain ponds or swamps. The moraine commonly has a steep north slope, but southward it merges with outwash plains and valley trains, which extend to the Susquehanna River. The Valley Heads moraine cannot be continuously traced across the uplands between through valleys, though it exists in many places as a belt of constructional topography with but low knolls, ridges and closed depressions.

North of the Valley Heads moraine the drift is thicker and the topography is more obviously a product of glaciation. Lithologic content of the till is more varied than south of the moraine, including pebbles of igneous and metamorphic rocks as well as abundant limestone and calcareous shale. The matrix is dark gray, well graded and contains free calcium carbonate. The matrix is increasingly calcareous northward from the Valley Heads moraine, reaching its maximum carbonate content in the northern portion of the transect near the exposure area of the Onondaga limestone.

A number of recessional moraines have been mapped north of the Valley Heads terminal position (Carney, 1909; Taylor, 1924, 1932). These moraines are believed to be of minor importance and the drift at the north end of the transect (fig. 1, line X-Y) is probably only slightly younger than the Valley Heads moraine.

SITE SELECTION

This study was designed primarily to test the hypothesis that depth of leaching of Wisconsin till in New York is a function of its carbonate content. Accordingly, an attempt was made to select sampling sites similar enough to minimize the effects of factors other than carbonate content.

To minimize slope differences, slopes were measured and sampling sites were limited to flat or gently sloping positions. Drainage differences were minimized by selection of sites only on moderately well drained and well drained soils (Soil Survey Staff, 1951, p. 169-172). To avoid large differences in permeability, sampling was restricted to sites with compact lodgment till (Flint, 1957, p. 120) and bulk-density measurements were made (table 2). The size of the carbonate-bearing particles and the texture of the lower horizons

of the profiles were measured. Rainfall and temperature do not vary enough in the study area to account for differences in depth of leaching (Mordoff, 1949).

All sites are located on uplands between major valleys where the drift is composed primarily of ground moraine and is probably more uniform from place to place than in the valley fill. Water-laid drift, glacial lake and stream deposits and ablation moraine were avoided as having greater horizontal and vertical variability. Because of the thinness of the upland drift, calcareous till was located in only very few sites in the southern part of the area.

Preliminary field studies were made in Schuyler, Tioga, Cortland, Tompkins and Cayuga counties (fig. 1). Samples were collected and profiles examined in a broad transect across the uplands east and southeast of Cayuga Lake, covering catenas of soils ranging from very low- to high-lime content (Cline, 1955) and ranging south across the Valley Heads moraine.

FIELD PROCEDURE

At each site selected along the north-south transect, depth of leaching was determined by testing with cold dilute hydrochloric acid, the soil profile was described and topographic position was recorded. Samples were taken for determination of carbonate content at one or more depths beneath the leached zone. One sample was taken at least 3 feet below the leached zone at each site where this was possible.

At a number of sites, selected to represent significantly different pedologic and geologic situations along the transect, samples were taken from the lower part of the B horizon and from the C horizon for mechanical analysis and bulk-density determinations.

Depth of leaching was determined at a large number of other sites. Although calcareous material was not reached at many of these sites, either because of thinness of the mantle, or difficulty of excavation, the depth examined afforded a minimum value exceeded by the depth of leaching.

Standard procedures and terminology of the Soil Survey Manual (Soil Survey Staff, 1951) were followed in soil-profile descriptions. Hydrogen-ion concentration was measured by means of bromcresol green, chlorphenol red and bromthymol blue indicators.

LABORATORY PROCEDURE

Carbonate content was determined separately for material less than 2 mm and for material 2 mm to 5 mm in diameter (table 1) following the procedure approved by the Association of Official Agricultural Chemists (1950) for analysis of liming materials. A weighed amount of the sample is boiled in excess hydrochloric acid, then back titrated with sodium hydroxide. Because no effort was made to distinguish between calcium and magnesium carbonates, the data are expressed in terms of calcium carbonate equivalence. For material from 5 mm to about 5 cm in diameter, lime content was estimated by comparing the weight of the calcareous pebbles with the total weight of the size fraction. No effort was made to distinguish between calcite and dolomite.

Mechanical analyses of the soil fraction less than 2 mm in diameter (table 2) were made by the pipette method described by Kilmer and Alexander

(1949). Preliminary experiments showed that the hydrogen peroxide treatment could be eliminated because of the very low organic content of the samples, all of which were taken at least 19 inches below the surface. Duplicate determinations were made for the first 8 samples to control analytical error. Thereafter, one duplicate determination was made in each set of 8. The weight percents of fragments from 2 mm to 5 mm in diameter and of fragments from 5 mm to about 5 cm in diameter were determined from the bulk samples.

Bulk density (table 2) was determined by the paraffined clod method modified after Shaw (1917). Measurements were made on three clods for each horizon. The air dry weights of the clods used ranged from 5 to 116 grams. After bulk density of the clod had been determined, material larger than 2 mm was separated by sieving and the bulk density was recalculated on the basis of the material smaller than 2 mm.

DEPTH OF LEACHING AS RELATED TO CARBONATE CONTENT

Estimates of carbonate content have commonly been based on the ratio of calcareous pebbles to total number of pebbles in a given sample. At the outset of the present investigation it was considered that differences in pebble size and in form and concentration of carbonate might make pebble counting an invalid substitute for quantitative determination of calcium-carbonate equivalence. Accordingly, carbonate contents of several size fractions were tested for representativeness and for information on the relation of carbonate distribution to particle size.

Sand, silt and clay are better suited than are particles coarser than 2 mm for statistically adequate sampling and application of quantitative chemical methods of carbonate determination. Table 1 shows that carbonate content of the fraction finer than 2 mm (column 3) corresponds very closely with total calcium-carbonate equivalent for the sample exclusive of cobbles and boulders (column 5).

A more rigorous test of the acordance of data based on different size fractions is illustrated by comparison of the calcium-carbonate equivalents of the 2 to 5 mm fraction with those of the fraction less than 2 mm in diameter. For high values of carbonate equivalence the ratios approach unity and interpretations based on components somewhat coarser or finer than 2 mm may be counted on to yield comparable results. In the range between 1 percent and 20 percent calcium carbonate equivalent of the fraction finer than 2 mm the coarser material is about half again as calcareous as the finer fraction. Below 1 percent the coarser fraction is several times as calcareous. Carbonate content of the coarser fraction becomes decreasingly representative as total carbonate decreases.

A crude approximation of the carbonate content of the coarser components was obtained by testing all particles between 5 mm and 5 cm in diameter with cold dilute hydrochloric acid. Pebbles determined to be calcareous by this test include not only limestone and dolostone but also limy shale and sandstone with low carbonate content. The range in carbonate equivalence for any given percent of calcareous pebbles may be great and becomes a proportionately greater source of error as the number of pebbles decreases. High

TABLE 1

Depth of Leaching in Relation to Calcium-carbonate Equivalent of Indicated Fractions of Lodgment Till Under Soils of Very Low- to High-lime Status

Site	Depth of leaching	$CaCO_3$ equivalent of the fraction			calcareous pebbles by weight 5 mm-5 cm	Soil	
		<2 mm	2-5 mm	<5 mm		Series	Lime status
	(in.)	%	%	%	%		
95	192	1.2	1.9	1.3	1.9	Mardin	Very low
79	90	0.1	1.1	0.3	0.0	Mardin	Very low
80	90	0.8	4.1	1.1	0.0	Bath	Very low
94	84	5.0	9.3	5.4	47.6	Bath	Very low
82	67	17.0	21.0	17.2	81.9	Lansing	Medium
99	60	5.9	9.9	6.2	77.8	Lansing	Medium
100	54	2.6	4.7	2.8	4.8	Mardin	Very low
78	54	2.8	3.7	2.9	0.0	Valois	Low
30	48	16.4	28.3	17.4	84.3	Conesus	Medium
32	45	11.7	22.8	12.8	19.6	Conesus	Medium
26	38	12.4	7.0	12.1	28.3	Lansing	Medium
83	37	17.9	25.3	18.3		Lansing	Medium
49	32	17.9	31.7	18.5	78.0	Lima	High
48	29	27.1	29.3	27.3	98.5	Honeoye	High
46	28	30.5	31.8	30.7	87.3	Honeoye	High
45	27	26.4	31.7	26.8	81.5	Honeoye	High
31	24	19.1	29.1	20.0	82.2	Honeoye	High
28	24	19.4	29.2	20.2	80.5	Honeoye	High
85	22	28.0	29.3	28.1	83.1	Lima	High
51	21	25.8	29.4	26.0	99.4	Honeoye	High

calcium-carbonate equivalent in the fraction finer than 5 mm is consistently accompanied by a dominance of calcareous pebbles in the coarse fraction (table 1). Where total carbonate is low, the volume percent of calcareous pebbles ranges widely. Under such conditions, pebbles must be recognized as an unreliable indicator of total carbonate content.

It is concluded that analysis of data from one rather than another of several possible size fractions smaller than pebble size may slightly alter measured carbonate values, but probably will not change comparative position or indicated relationships. As a matter of convenience the combined size fraction finer than 2 mm is employed in analyses and discussions to follow.

Data in table 1, plotted in figure 2, show the relationship between depth of leaching and the calcium-carbonate equivalent of underlying drift. The relationship holds alike for the fraction less than 2 mm in diameter and for that between 2 and 5 mm. Regression equations and correlation coefficients listed in figure 2 show the relationship to be highly significant. It will be observed that calcium-carbonate equivalents as plotted in this chart cluster into three distinct groups.

A key to the distinct clustering of data of figure 2 may be found in the Soil Association Map of New York (Cline, 1955) which shows areas of low-, medium- and high-lime soils. Low-lime soils are described as strongly acid to

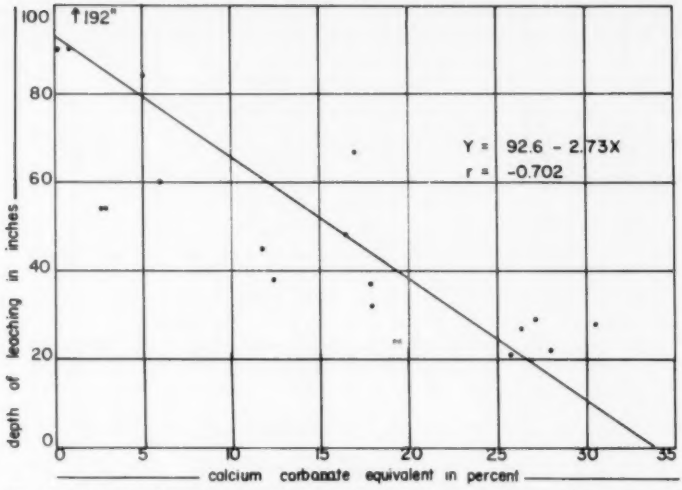

Fig. 2. Relationship between depth of leaching and calcium-carbonate equivalent of the fraction of lodgment till less than 2 mm in diameter.

depths greater than 30 inches but neutral or containing free carbonate deep in the substratum. Medium-lime soils have strongly acid upper parts to depths of more than 12 inches but lime is commonly found at depths below 30 or 40 inches. High-lime soils generally have free lime at 18 to 30 inches below the surface. Boundaries of the three areas as indicated on figure 1 are modified after Cline (1955).

In the area of low-lime till, soils of the Valois catena predominate. The boundary between low- and medium-lime soils (and east of Owasco Lake, between low- and high-lime tills) marks a major change in soil from acid soils with fragipans and belonging to the Valois catena to medium-lime soils with textural B horizons belonging to the Lansing catena. This boundary is not marked by topographic feature, but rather is a zone of rapid transition in soil characteristics ranging from ¼ to 1 mile wide. The area of medium-lime soils is a broad zone of gradual change in acidity of the solum, thickness of the A_2 horizon and depth to carbonates decreasing northward toward the boundary with high-lime till on which Honeoye and Lima soils predominate.

Association of the three-fold clustering of data in figure 2 with the low-, medium- and high-lime soils is confirmed in table 2 which shows mean values of leaching depth and calcium carbonate equivalence for each of the three groupings to be representative of the Valois, Lansing and Honeoye catenas, respectively.

Depth of leaching and carbonate content of the size fractions finer than 2 mm and 5 mm to 5 cm are plotted against site position in figure 3. Boundaries between areas of high-, medium-, and low-lime till are indicated where they intercept the transect line X-Y of figure 1. Sample sites are located on figure 3 by projection to X-Y, maintaining proportionate distances to soil

TABLE 2

Bulk Density, Particle-size Distribution and Calcium Carbonate Equivalent of Lodgment Till in Relation to Depth of Leaching at 9 Sites Representative of the Range of Soils Studied.

Site	Soil series	Depth of leaching, inches	$CaCO_3$ equiv. <2 mm	Bulk density of basal till		Particle size distribution			
				Whole soil	<2 mm	Sand	Silt	Clay	Silt & clay
Low & Very Low Lime Soils			%	gm/cc	gm/cc	%	%	%	%
95	Mardin	192	1.2	1.9 ± 0.1	1.7 ± 0.1	43.5	41.4	15.1	56.5
94	Bath	84	5.0	1.9 ± 0.1	1.7 ± 0.0	22.7	53.0	24.3	77.3
100	Mardin	54	2.6	2.1 ± 0.1	1.9 ± 0.0	24	51	25	76
78	Valois	54	2.8	1.8 ± 0.2	1.5 ± 0.1	28.5	46.7	24.8	71.5
Medium-Lime Soils									
82	Lansing	67	17.0	2.1 ± 0.1	2.0 ± 0.0	28.1	54.6	17.3	71.9
83	Lansing	37	17.9	1.7 ± 0.1	1.7 ± 0.1	36.7	48.6	14.7	63.3
High-Lime Soils									
49	Lima	32	17.9	1.7 ± 0.1	1.6 ± 0.1	32.9	48.9	18.2	67.1
46	Honeoye	28	30.5	2.1 ± 0.2	1.8 ± 0.5	27.7	54.9	17.4	72.3
45	Honeoye	27	26.4	1.9 ± 0.0	1.8 ± 0.0	40.9	49.3	9.8	59.1

parent material boundaries. Similarly indicated are exposure areas of the Onondaga and Tully limestones, the southernmost sources of carbonate in the till.

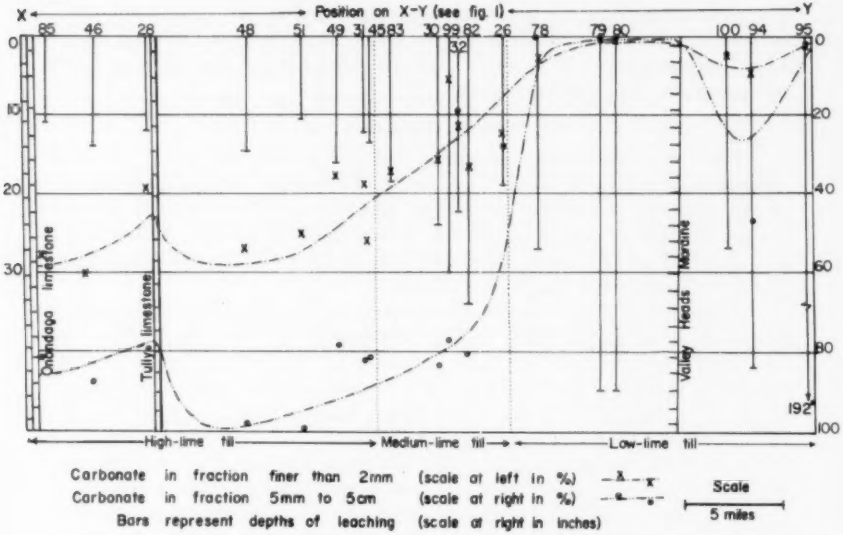

Fig. 3. Depth of leaching and calcium-carbonate equivalent of the fractions of lodgment till less than 2 mm and 5 mm to 5 cm in diameter, plotted in relation to position of sample sites with respect to Transect X-Y (see fig. 1).

No clear-cut trend in change of carbonate content is indicated between the areas of exposure of the Onondaga and Tully limestones, but from the Tully outcrop southward, calcium carbonate decreases and depths of leaching increase through the length of the transect. Across the area of high-lime till the graph shows a scattering of points, suggesting relatively small changes. Across the area of medium-lime till, the changes are rapid with carbonate content asymptotically approaching zero. Low carbonate values in the area of low-lime till and south of the Valley Heads moraine are accompanied by leaching depths of 6 to 7 feet. Station 95 with carbonates leached to 192 inches is considered unrepresentative and the south end of the transect shows no marked increase in depth of leaching at the Valley Heads border.

The pattern of change in carbonate content appears related to distance from source rocks, to avenues of principal ice movement and to intensity of erosion. The concentration of any given constituent in the basal load of a glacier may be expected to be depleted by deposition at a decreasing rate with increasing distance from the source. It may be conjectured that this is the principal process affecting change in carbonate content in the area of high-lime till. More erratic and rapid changes in concentration of a constituent in the basal load may result from rapid erosion which causes dilution of all previously derived components. Uptake of clastic rocks in the region of medium-lime till may account for the rather rapid decrease in carbonate content. Sup-

port for this hypothesis is found in the sharp drop in percentage of carbonate pebbles in this belt, for contamination of the basal load by clastic rock fragments would be most marked in the very coarse fraction.

It is clear from the preceding discussion that in the area of this study, carbonate content and depth of leaching vary oppositely and in such manner as to indicate marked dependence of depth of leaching on the initial carbonate content of the soil parent material.

LEACHING IN RELATION TO AGE

The till is necessarily somewhat younger in the northern part of the transect than it is in the south. This fact might be expected to account for part of the observed southward increase in depth of leaching. If zones of rapid change in depth to carbonate coincide with boundaries marking differences in age of till, it must be concluded that both duration of exposure and initial carbonate content are controlling factors.

A number of minor recessional moraines (Carney, 1909) indicate marginal positions occupied briefly during withdrawal of Wisconsin ice from the Allegheny Plateau. None of these moraines is strongly developed and, for reasons to be shown, the time interval represented in the area of this transect by till deposition north of the Valley Heads moraine is believed to have been no more than a matter of centuries.

The Valley Heads moraine on the other hand suggests an important still-stand prior to deposition of the minor recessional moraines. The Valley Heads border appears to lie well within the zone of low calcium carbonate content and deep leaching as shown in figure 3. Although the number of sample sites in low-lime till is limited, low hydrogen ion concentration and lack of carbonates here demonstrated in 36 additional sites where bedrock or stony substrata prevented sampling of calcareous till. The evidence indicates that in the uplands of the low-lime till zone, carbonates are generally leached to depths of 5 to 8 feet, but it does not demonstrate an increase in depth of leaching across the Valley Heads border. Thus the topographic feature most suggestive of time lapse during deglaciation lies wholly within the zone of low-lime till, very acid soils and considerable depth to carbonate.

For this reason, and because the most striking change in depth of leaching occurs in the zone of medium-lime till where no feature suggests even minor interruption of glacial recession, changes in carbonate content in this study area cannot be ascribed to different episodes of till deposition.

To avoid unduly minimizing age as a potential factor affecting the depth of leaching, it is necessary to determine as closely as possible the interval between deposition at opposite ends of the transect. Unfortunately, correlation of drifts of Wisconsin age in central New York with those in the Upper Mississippi basin is still somewhat tentative.

Two radiocarbon dates show the Valley Heads moraine to be no younger than Late Cary. Wood from the base of marly silt overlying gravel of the Chaffee outwash plain in the southeastern corner of Erie County, New York, must slightly postdate withdrawal of Valley Heads ice from its maximum position. This material was dated at 12,020 $\pm$ 300 B.P. (W-507). Organic material associated with mastodon remains preserved in a shallow basin north

of King Ferry (between sites 48 & 51, fig. 1) could not have been deposited until deglaciation of the area of this study. This material was dated at 11,410 ± 410 (Y-460). These data confirm the existing interpretation that the southern New York section of the Allegheny Plateau was completely deglaciated during and since Two Creeks time (Flint, 1956).

Two posibilities must be acknowledged for the age of till south of the Valley Heads moraine, and therefore also for the time range represented by surficial till in the study area. Minimum range is justified by acceptance of MacClintock and Apfel's (1944) correlation of drift in the Chenango Valley with the Wisconsin terminal border west of the Salamanca re-entrant in western New York. This border, marked by Binghamton moraine of MacClintock and Apfel, has been correlated with the Kent moraine of northeastern Ohio and northwestern Pennsylvania by continuous tracing (Muller, 1957b; White, et al., 1957). Minimum age of the Kent moraine is given by radiocarbon dating of marl and peat from a bog at Corry in Erie County, Pennsylvania. Deposition in this kettle 6 miles inside the Wisconsin drift border presumably commenced shortly after the beginning of deglaciation. Radiocarbon assay of 14,000 ± 350 years B.P. on basal marl (W-365) from the Corry bog (Rubin, et al., 1958) suggests that the Kent moraine was deposited near the beginning of the Cary substage. The interval of till deposition in the area of the present study could then be no more than about 2000 years, ending 12,000 years ago. On the basis of these correlations, the lack of marked contrast in depth of leaching across the Valley Heads moraine might have been predictable.

In questioning the presence of Binghamton drift south of the Valley Heads moraine in the Finger Lakes region, Denny (1956) opened up the possibility of greater age for the oldest till in the present study area. If the drift south of the Valley Heads moraine is related to the Olean drift of MacClintock and Apfel (1944) it is of probable Tazewell correlation (Flint, 1953; MacClintock, 1954a), with possible age of 17,000 to 19,000 years before the present. The interval of till deposition in the area of this study would then be 5000 to 7000 years, ending 12,000 years ago. The ratio of age of the youngest till in the study area to difference in age between oldest and youngest till would then be 2:1. Under such conditions age difference might have been expected to produce a contrast in depth of leaching across the Valley Heads border. Nonexistence of such a contrast may be interpreted either as evidence against the importance of age as a control on depth of leaching, or as evidence of post-Tazewell age for drift at the southern end of the study area.

LEACHING IN RELATION TO OTHER FACTORS

Factors other than carbonate content and age were held reasonably constant by sampling. Several factors which could affect results of the foregoing investigation are slope, particle size distribution, permeability and subsequent burial or erosion of the drift surface.

Slope differences are too small to be considered significant. Particle-size variation is considerable among the sites, but it is seemingly random with respect to calcium-carbonate equivalent, so that there is no evidence that it consistently modifies the dominant relationship of calcium carbonate and depth of leaching. Consistently high bulk-density measurements in unweathered till

across the transect suggest rather uniformly slow rates of water movement in all soils studied. No consistent relationship of bulk density to depth of leaching was recognized.

Erosional truncation of the soil profile is aparently negligible in the case of all but one of the sites sampled. At site 28, slight erosion was apparent and the measured depth of leaching is somewhat less than might be expected on the basis of carbonate content alone. No evidence of postglacial burial was found at any of the sites sampled. At only one site is there stratified material within the leached zone. At this location, site 79, stratified material occurs 4 feet below the surface. The greater permeability of this zone might be expected to accelerate leaching, but the relationship between lime content and depth of leaching at this site is consistent with other data in the transect.

CONCLUSIONS

Within soils in till of comparable age, density and texture on similar slopes in central New York, depth of leaching is primarily a function of original carbonate content. The relationship of depth of leaching to calcium-carbonate equivalent should apply to other areas where additional contributing factors range within sufficiently narrow limits.

In the area of this study, depth of leaching increases and carbonate content decreases consistently southward from the Tully limestone exposure area. Both depth of leaching and carbonate content change most rapidly with distance in an area lacking evidence of stillstand during glacial recession. This pattern of change is attributed to depletion of carbonate in the basal load of the glacier and dilution by entrainment of clastic material. Under the control of initial carbonate content, depth of leaching varies as much within a single drift as it does across a drift border.

Depth of leaching alone, without knowledge of variation of carbonate content, may prove an unreliable criterion of relative age of drift sheets. Where total carbonate content is low, pebble counts may prove an unreliable indicator of carbonate equivalence.

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REFERENCES

Alden, W. C. and Leighton, M. M., 1917, The Iowan drift: Iowa Geol. Survey Ann. Rpt., v. 26, p. 49-212.

Allen, B. L. and Whiteside, E. P., 1954, The characteristics of some soils on tills of Cary and Mankato age in Michigan: Soil Sci. Soc. America Proc., v. 18, p. 203-206.

Association of Official Agricultural Chemists, 1950, Methods of analysis: Washington.

Bradfield, R., 1941, Calcium in the soil—I: Soil Sci. Soc. America Proc., v. 6, p. 8-15.

Bryan, K. and Albritton, C. C. Jr., 1943, Soil phenomena as evidence of climatic changes: Am. Jour. Sci., v. 241, p. 469-490.

Carney, F., 1909, The Pleistocene geology of the Moravia Quadrangle, New York: Denison Univ. Sci. Lab. Bull., v. 14, p. 335-342.
Chandler, R. F., Jr., 1941, The amount and mineral nutrient content of freshly fallen leaf litter in the hardwood forests of central New York: Am. Soc. Agronomy Jour., v. 10, p. 859-871.
Cline, M. G., 1953, Major kinds of profiles and their relationships in New York: Soil Sci. Soc. America Proc., v. 17, p. 123-127.
Cline, M. G., 1955, Soils and soil associations of New York. Cornell Ext. Bull. 930, 72 p.
Conrey, G. W., Paschall, A. H., and Burrage, E. M., 1937, A key to the soils of Ohio: Ohio Agr. Expt. Sta. Spec. Circ. (Rev.), 44.
Denny, C. S., 1956, Wisconsin drifts in the Elmira region, New York, and their possible equivalents in New England: Am. Jour. Sci., v. 254, p. 82-95.
Dreimanis, A., 1957, Depths of leaching in glacial deposits: Science, v. 126, p. 403-404.
Ehrlich, W. A. and Rice, H. M., 1955, Postglacial weathering of Mankato till in Manitoba: Jour. Geology, v. 63, p. 527-537.
Ehrlich, W. A., Rice, H. M. and Ellis, J. H., 1955, Influence of the composition of parent materials on soil formation in Manitoba: Canadian Jour. Agr. Sci., v. 35, p. 407-421.
Ellis, J. H., 1938, The soils of Manitoba: Winnipeg, Manitoba Econ. Survey Board.
Fairchild, H. L., 1932, New York moraines: Geol. Soc. America Bull., v. 43, p. 627-662.
Flint, R. F., 1949, Leaching of carbonates in glacial drift and loess as a basis for age correlation: Jour. Geology, v. 57, p. 297-303.
————, 1953, Probable Wisconsin substages and late-Wisconsin events in northeastern United States and southeastern Canada: Geol. Soc. America Bull., v. 64, p. 897-920.
————, 1956, New radiocarbon dates and late Pleistocene stratigraphy: Am. Jour. Sci., v. 254, p. 265-287.
————, 1957, Glacial and Pleistocene geology: New York, John Wiley and Sons, 509 p.
Holmes, C. D., 1935, Glacial and interglacial development of Chittenango Falls State Park: Am. Jour. Sci., v. 29, p. 41-47.
Holmes, C. D., 1952, Drift dispersion in west-central New York: Geol. Soc. America Bull., v. 63, p. 993-1010.
Howard, A. D., 1946, Caliche in glacial chronology (Abstr.): Geol. Soc. America Bull., v. 57, p. 1204.
Jenny, H. and Leonard, C. D., 1934, Functional relationships between soil properties and rainfall: Soil. Sci., v. 38, p. 363-381.
Kay, G. F., 1931, Classification and duration of the Pleistocene: Geol. Soc. America Bull., v. 42, p. 425-466.
Kay, G. F. and Graham, J. B., 1943, The Illinoian and post-Illinoian Pleistocene geology of Iowa: Iowa Geol. Survey, v. 38, 262 p.
Kilmer, V. J. and Alexander, L. T., 1949, Methods of making mechanical analysis of soils: Soil. Sci., v. 68, p. 15-24.
Leighton, M. M., 1923, The differentiation of the drift sheets in northwestern Illinois: Jour. Geology, v. 31, p. 265-281.
Leighton, M. M. and MacClintock, P., 1930, Weathered zones of the drift sheets of Illinois: Jour. Geology, v. 38, p. 28-53.
MacClintock, Paul, 1954a, Leaching of Wisconsin glacial gravels in eastern North America: Geol. Soc. America Bull., v. 65, p. 369-383.
————, 1954b, Pleistocene geology of the St. Lawrence Lowland: N. Y. State Sci. Service, Rept. Inv. 10.
MacClintock, Paul and Apfel, E. T., 1944, Correlation of the drifts of the Salamanca reentrant, New York: Geol. Soc. America Bull., v. 55, p. 1143-1164.
McCaleb, S. B., 1954, Mineralogical properties of the Gray-brown Podzolic-Brown Podzolic soil sequence: Soil Sci., v. 77, p. 319-333.
McCaleb, S. B. and Cline, M. G., 1950, Physical and chemical properties of Brown Forest and Gray-brown Podzolic soils: Soil. Sci., v. 70, p. 315-328.
Mordoff, R. A., 1949, The climate of New York State: Cornell Ext. Bull. 764.
Muller, E. H., 1957a, Filled bedrock gorges in the drainage basin of Cayuga Lake, New York (Abstr.): Geol. Soc. America Bull., v. 68, p. 1771.
————, 1957b, Glacial geology of western and central New York (Abstr.): Geol. Soc. America Bull., v. 68, p. 1897.
Rubin, Meyer and Alexander, Corinne, 1958, U. S. Geological Survey Radiocarbon dates —IV: Science, v. 127, p. 1476-1487.
Russell, J. C., and Engle, E. G., 1925, Soil horizons in the central prairies: Am. Soil Survey Assoc., Rept. 5th meeting, v. 6, p. 1-18.

Schultz, C. G., Leuninghoener, G. C. and Frankforter, W. D., 1951, Graphic resume of the Pleistocene of Nebraska: Univ. Nebraska State Mus. Bull., v. 3, p. 6.

Shaw, C. F., 1917, A method for determining the volume weight of soils in field condition: Am. Soc. Agronomy Jour., v. 9, p. 38-42.

Soil Survey Staff, 1951, Soil survey manual: U. S. Dept. Agriculture Handbook, v. 18.

Tarr, R. S., 1909, in Williams, H. S., Description of the Watkins Glen-Catatonk district, New York: U. S. Geol. Survey, Geol. Atlas, Folio 169.

Taylor, F. B., 1924, Moraines of the St. Lawrence Valley: Jour. Geology, v. 32, p. 641-667.

Taylor, F. B., 1932, Recessional moraines in the Buffalo-Genesee and Finger Lakes areas (Abstr.): Geol. Soc. America Bull., v. 43, p. 189-190.

Thornbury, W. D., 1940, Weathered zones and glacial chronology in southern Indiana: Jour. Geology, v. 48, p. 449-475.

von Engeln, O. D., 1921, The Tully glacial series: New York State Mus. Bull., v. 227-228, p. 39-62.

———, 1929, Interglacial deposit in central New York: Geol. Soc. America Bull., v. 40, p. 469-479.

———, 1931, A preglacial or interglacial gorge near Seneca Lake, N. Y.: N. Y. State Mus. Bull., v. 286, p. 127-131.

———, 1938, Large, sharply defined terminal moraine ridges: Internat. Geog. Union Cong. Amsterdam, Comptes Rendus, v. 2, p. 211-213.

White, G. W., Droste, J. B., Sitler, R. E. and Shepps, V. C., 1957, Glacial deposits of northwestern Pennsylvania (Abstr.): Geol. Soc. America Bull., v. 68, p. 1902-1903.

Willman, H. B. and Payne, J. N., 1942, Geology and mineral resources of the Marseilles, Ottawa and Streator quadrangles, Illinois: Ill. Geol. Survey Bull. 66.

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AN INVESTIGATION INTO PROCESSES OCCURRING IN SOLIFLUCTION

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ABSTRACT. The mechanics of solifluction downslope of a late snow patch has been investigated using instrumental techniques new in this field, for the determination of soil pore-water pressures, density and movement. Substantial changes in water content and considerable sub-surface water flow occur together with changes in state of stress. Low and sub-atmospheric pore-water pressures were observed. To explain solifluction as being analogous to soil movement induced by impeded drainage over an impermeable layer is an over-simplification, which ignores fundamental characteristics of newly-thawed and frozen soils. It is doubtful to what extent concepts of soil strength applicable under normal conditions apply during freeze-thaw processes.

INTRODUCTION AND ACKNOWLEDGEMENTS

Preliminary investigations of solifluction features were made in 1953, 1954 and 1955 in central Norway. These involved several new techniques and repeated observations (Williams 1957a, 1957b). The importance of detailed instrumental studies of processes in solifluction was obvious. Accordingly, during five weeks in June, July and late August, 1956 a team with an average strength of five maintained an instrument layout and recorded readings and observations on solifluction below a late snow patch. A cyclostyled report (Williams, et al., 1957) detailed these methods, and stimulated discussion. Subsequently, further field observations were made in March, June and August, 1957.

The field work was part of a program supported by the Royal Geographical Society, Norwegian Nansen Fund and others. Willing Cambridge and Oslo students, of whom Mrs. Kari Williams was the most persistent, were field assistants during the work. Mr. J. A. Pihlainen, National Research Council, Canada, carried out the statistical analyses. To these, thanks are extended.

SITUATION AND GENERAL DESCRIPTION

The occurrence of solifluction investigated (pl. 1), on the bank of the Einøvling stream near Svaanalegret, about 8 kilometers southeast of Snøhetta, is similar to many others in Dovrefjell, Trollheimen, Rondane, and elsewhere above the tree line. Permafrost is absent and the mean annual temperature is −0.9°C (Det Norske Meteorologiske Institutt 1949-1955). In this example, the snow patch persists only on the upper part of the 30-meter high valley side. Solifluction of the same type occurs for some 400 meters along the valley side and there is a conspicuous break of slope about two-thirds of the way up (see pl. 1 and fig. 1). This break of slope might be interpreted as remains of a glacio-fluvial terrace of the type commonly found in this area (Strøm, 1952). However, the rate of soil movement is great enough at present that it alone could be sufficient to produce the break of slope. The soil is till, with many stones and boulders, clay and sand lenses, in a silty-clay matrix. Depth to bedrock is not known but is greater than 2 meters.

Six resistance thermometers consisting of glass-encased thermistors built into simple probes of 1 to 2 meters length were placed at various depths in

PLATE 1

A. Site of the solifluction investigations described. (photo: D.R.M. Lillistone)

B. The leads from the thermistors and soil movement probes were run to the tent seen at right center left. (photo: Kari Williams)

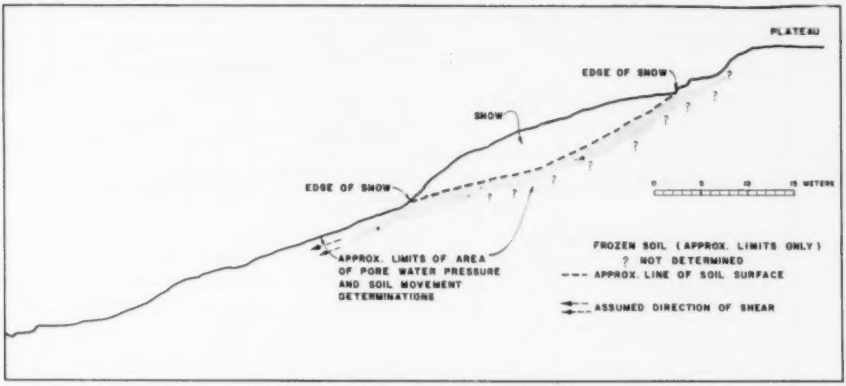
PLATEAU
EDGE OF SNOW
SNOW
EDGE OF SNOW
APPROX. LIMITS OF AREA OF PORE WATER PRESSURE AND SOIL MOVEMENT DETERMINATIONS
FROZEN SOIL (APPROX. LIMITS ONLY)
? NOT DETERMINED
APPROX. LINE OF SOIL SURFACE
ASSUMED DIRECTION OF SHEAR

Fig. 1. Section through slope showing site investigated, late June condition. The external soil and snow surface is a true profile surveyed by Abney level, about June 20, 1957. The other data are based on various observations in 1956-1957.

holes drilled in the still-frozen slope (fig. 2). Eighteen mercury soil temperature thermometers were placed at various positions up to 40 centimeters deep in the soil. Results of the temperature recordings were limited but together with some probing, the general shape of the still-frozen soil mass was established (see fig. 1). The temperature of the still-frozen layer was virtually constant throughout (the thermistor probe sensitivity being about $\pm$ 0.05°C).

The snow accumulates in the winter because of wind blowing it from the plateau at the top of the slope. This plateau was observed to be free of snow in March, and the presence of stony earth circles—a form of patterned ground developing only where there is no snow cover—(Williams, in press) verify its general absence there.

INVESTIGATIONS INTO FACTORS INFLUENCING SOIL STABILITY

Theoretical background.—General explanations of solifluction involving concepts such as "lubrication of soil particles by water" are incomplete and often misleading. Terzaghi (1950) citing Hardy (1919) points out that water acting on quartz is not a lubricant. The study of processes in solifluction phenomena is a study of soil mechanics, and such special conditions are met with under freezing and thawing as to demand investigation of many basic soil properties.

According to soil mechanics theory, movement in soil occurs when the forces tending to produce shear (shear stress) exceed the strength of the soil (its resistance to shear). The latter is assumed to consist of two components, that due to internal friction and that due to cohesion. Cohesion in soils is ascribed severally to the properties of the adsorbed water in close proximity with the soil grains, the mutual attraction of particles, a grain structure inherent in clay soils, apparent cohesion which is due to the capillary forces in the pore-water in unsaturated soil, etc. According to the work of Bjerrum (1954), based on that of Hvorslev (1937) and others, in saturated soils the

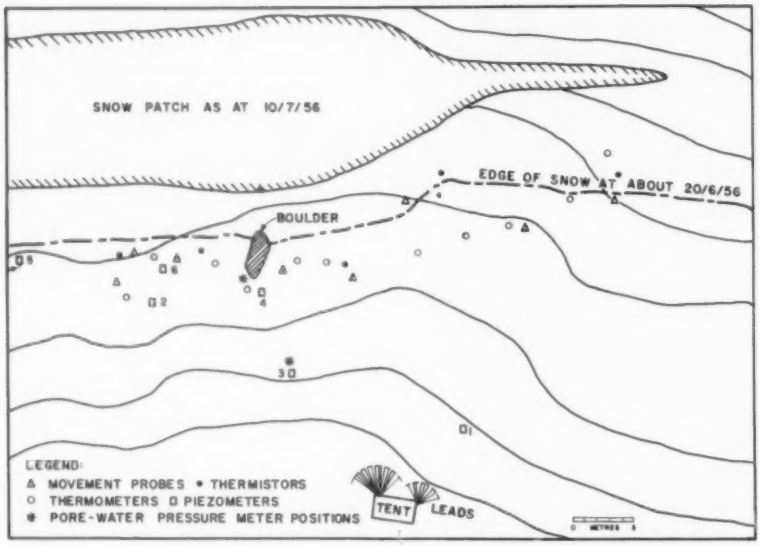

Fig. 2. Map of instrument layout on solifluction slope below late-lying snow (drawn from compass survey and field sketches). Contour interval—2 m.

magnitude of cohesion is dependent on water content and therefore on void ratio $= \left(\frac{\text{vol. voids}}{\text{vol. solid}}\right)$ The internal friction developed depends on the pressure at the intergranular contacts, $\bar{\sigma}$, which varies with U_w, the pore-water pressure, according to the relations $\bar{\sigma} = \sigma - U_w$ where σ is the total weight of wet soil above. Pore-water pressure, therefore, also directly affects soil strength.

Soil movement.—Soil movement was recorded directly by the use of special electrical probes described fully elsewhere (Williams, 1957b). Movement was also apparent through visible distortion of the probes and thermistor probes. In any one year, soil movement on the slope is largely restricted to several patches of a few square meters in area, although a large part of the slope is in an unstable condition. Maximum surface movement during the spring of 1956 was about 25 centimeters downslope. It is not possible to distinguish those parts actively moving, except by repeated observations. The movements occur at progressively greater depth as the season advances, and to a depth of at least 75 centimeters. At any given position, movements occur gradually, taking place over several weeks during the spring. Other than upslope from the snow patch, no distinct shear planes were visible. The movements recorded showed no regular or periodic variations of a diurnal or similar nature.

Effect of snow patch.—The late-lying snow patch in summer to some extent loads the slope, and the possibility of this increasing the shear stress must be considered. The initial solifluction movements will tend to remove this in-

crement of the shear stress from the soil downslope of the snow (see fig. 1). It will only persist or be reapplied if there is also downslope movement of the still-frozen soil, which is unlikely. In other words, the weight of the snow patch is probably supported almost entirely by the still-frozen soil at least after the initial soil movements have occurred and has, therefore, little significance in causing movement. The movement is also substantially independent of the size of the snow patch which is steadily decreasing during thaw. The effects of the snow in causing movement are probably indirect.

Pore-water pressure determinations.—Pore-water pressure was determined at locations in the slope (fig. 2) with piezometers using simple mercury manometers attached to lengths of brass tubing with an internal diameter of about 7 mm, whose lower ends were at a depth of 50 to 115 cm. The brass tube was pushed into the thawed soil (in two cases, partially into holes drilled into still-frozen soil) while a rod was held in position inside the tube to prevent the entry of mud. Before attaching the filled manometer, the tube was filled with boiled water. Difficulty was experienced because of the rapid drainage from the tube into the soil. An electrical apparatus was also used, and confirmed the general nature of the pore-water pressures, although being far less sensitive.

Since water could be seen on the soil surface around most of the tubes and the soil was apparently saturated, it would be expected that the pressure of water (in centimeters of water) in the soil at any depth would equal approximately the depth to which the tubes were inserted, at least at those points. If complete drainage of excess water (above saturation point) did not occur, it might be greater than this. This latter condition (hydrostatic excess pressure in soil mechanics terminology) follows the transfer of some part of the weight of mineral soil to the entrapped pore-water. As explained above the greater the pore-water pressure, the smaller the component of soil strength due to internal friction. The pressures found, however, were considerably lower than expected, varying between −103 cm of water at 90 cm depth (sub-atmospheric pressures being expressed as negative) to +41 cm of water at 44 cm depth. About 250 readings, well distributed between these extremes, were taken at seven points, at intervals of half to several hours for periods up to three weeks. The lowest pressures (i.e. with the numerically highest, negative value) were recorded from depths in the proximity of the boundary between the thawed and frozen soil; surface water was also visible at these locations. Rhythmic diurnal changes of about 10 cm of water pressure were commonly observed. A gradual trend towards a minimum pressure and a subsequent rise was observed at most points during the three weeks, the total range at a single point being about 30 to 60 cm (see fig. 3; air temperatures for the period are shown in Williams, 1957b, fig. 3). Several more weeks elapsed before the soil had reached its driest summer condition. There was no clear relationship of pressure variations to rainfall.

These unexpected observations are of considerable significance and their causes are discussed below. The resulting effects are not easily interpreted. In general, lower pressures give greater strength by increasing internal friction, but even larger pore-water tensions are produced by drying in the summer

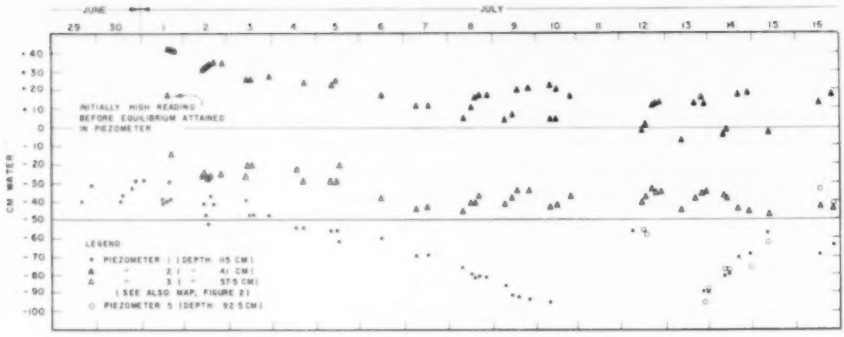

Fig. 3. Pore-water pressures at four locations in the slope.

(see Aitchison, 1957), so that the spring condition is to be interpreted as one of *relatively high* pore-water pressures.

It is often tacitly assumed that, in thawing of soil where considerable water accumulation has occurred at freezing, conditions of hydrostatic excess pressure, or at least a significant positive pressure, will arise generally and contribute in considerable measure to the instability. This is shown not to be the case. It is particularly tempting to believe that spontaneous liquefaction might occur (Terzaghi, 1950) where there is a sudden collapse or settlement of the soil grains with a transfer of a considerable part of their weight to the pore-water. However, failure to detect such high pore-water pressures in this slope indicates this to be of little significance in causing movement.

Frost heave.—Soils within the fine-sand-clay range, provided that sufficient water is available, show expansion at freezing several times greater than could be explained on the basis of the conversion to ice of the initial water content. The subject of much recent research, this expansion is due to accumulation of water during freezing, producing discrete ice layers, or individual crystals and masses which are often, but not always, visible to the naked eye (see for example, Beskow, 1935; Johnson, 1952). Such accumulation occurs as long as there is a sufficient supply of water available in the soil. The exact relation of frost heave to water availability is not fully understood but substantial heave occurs in the zone of capillary saturation. Recent work shows that heave may be caused by migration of water in soils that are only partly saturated (Penner, 1956). Most theories of frost heave attach great significance to the observed depression of the freezing point in soils which is generally supposed to be related to a considerable extent to effects of the surface forces of the particles, and hence to pore size.

Expansion per unit volume of the soil on this slope was calculated from density (gms. wet soil per cc) determined on soil samples from a depth of 5 to 20 cm. Using a modified form of the sand replacement method (West, 1953) determinations were made in the field in the spring (still-frozen and newly-thawed condition) and late summer, spread over two week periods. Subsequently, water contents were determined in the laboratory, permitting calculation of *in situ* dry densities (grams of mineral and organic soil per cc).

The ratio of average dry density for the spring series to that for the late summer series gives the contraction occurring during the summer, and is hence a measure of expansion due to frost heave.

Since the soil is heterogeneous, analysis of a number of samples from both periods is necessary. A maximum error of 6 percent is assumed possible in a single determination. The observations are summarized in the following table:

	No. of Determinations	Average Wet Density (grams wet soil per cc)	Average Dry Density (grams dry soil per cc)	Average Water Content (grams water per cc wet soil)
Year: 1956 Spring	14	1.74	1.37	0.37*
Late Summer	13	1.74	1.46	0.28
Year: 1957 Spring	10	1.75	1.30	0.45*
Late Summer	7	1.77	1.52	0.25

* The low water contents result from selecting some samples from the *thawed* soil, which had to be sufficiently drained to permit use of the sand replacement method.

The following conclusions can be drawn:

(a) Frost heave occurs in this slope. This is shown by the significantly different average values for spring and late summer dry densities and by the occasional observation of ice layers in the frozen soil. The slope is rather dry in late summer but the incomplete drainage due to a frozen layer at some centimeters depth during autumn freeze-thaw cycles, of rain or melt water liberated from early snow on the slope, may provide a source of water for the ice formation and frost heave.

(b) It is also established statistically that the heaved condition persists into the newly thawed state. A significance test on the difference between dry density in the spring (thawed soils only) and late summer of 1957, was based on the following data:

	Spring	Late Summer
M = samples =	7	7 (total samples taken)
$\bar{x}$ (average)	1.32	1.52
S (sample variance)	0.0451	0.0188

The number of samples, however, is insufficient to derive a value for the percentage expansion with any certainty.

(c) The *in situ* wet densities (unit volume weight) for spring and late summer periods are very similar. It is often stated that increased water content increases the likelihood of movement, because its weight adds to the shearing stress. In this case of heaved soil, there is little general increase in soil weight during the thaw period.

(d) In spite of the high water content, air filled some of the voids.[1] This was shown by specific weight determinations made on solid material from many of the frozen samples, enabling calculation of volume of voids per unit volume of soil. The latter is compared with water (ice) content.

An increase in the void ratio has several results in the thawed soil. Permeability is increased substantially; for example, Taylor (1948, p. 116) cites a 15 percent increase in void ratio of a cohesive soil, which increased permeability by ten times. Where ice-layer formation has occurred, the increase in permeability is likely to be primarily in a direction parallel to the ice layers, and hence parallel to the surface. It is not possible to get a quantitative value of permeability changes from calculations of void ratio in this case.

EFFECTS OF FROST HEAVE ON PORE-WATER PRESSURE AT THAW

That this increased permeability persists after thawing (although to an unknown degree) and is effective in a direction likely to be followed by melt water flow, partly explains the unexpectedly low pore-water pressures observed in this slope. Consolidation of the soil (the packing closer of grains) and liberation of water by thawing takes place only slowly and drainage of the water occurs sufficiently rapidly for pore-water pressures to remain low. Entry of air or surface water which would dissipate negative pressures is thought to be prevented because of the low permeability assumed to exist normal to the surface and the partial closure of capillaries by the snow mass or frozen soil higher up the slope.

A further explanation is supported by the work of Nersessova, as described by Tsytovich (1957). In her laboratory experiments, she found thawing of soil takes place progressively as the temperature approaches 0°C. Water is liberated according to temperature and the manner in which it is held in the soil, the most firmly held (bound) water thawing at the lowest temperature. The percentage of unfrozen water at a given negative temperature also varies with grain size composition.

Obviously the most firmly held water is that least likely to drain, but according to Tsytovich (1947) migrations of unfrozen water in frozen soil occur. The writer suggests that during the several weeks when the frozen layer in this slope has a temperature approaching 0°C it has a definite permeability, so that some drainage occurs over a long period prior to complete thaw. This will tend to lower the pressure in the water. It is well known that sub-atmospheric pore-water pressures are developed in the soil below the frozen layer, during freezing (see Penner, 1956). Water may then be drawn down from the frozen layer during partial thawing. The pressures may be further lowered by the volume decrease of the water (ice) taking place on the melting of some of the ice, while the expanded (heaved condition) of the soil is maintained by the longest remaining ice structures. These effects are likely to be accentuated by the heterogeneous nature of the soil and the range of pore sizes present. The existence of sub-atmospheric pressures in the unfrozen water of frozen soils was pointed out by Lovell (1957), and it is probable that these persist from the time when freezing occurred. To what extent the low pore-

[1] The term "void" includes all pores filled either with water or air.

water pressures observed should be related to each of these possible causes is not clear.

During the most active thaw period, therefore, the soil is in a unique condition. The water content is extremely high, and far above the normal saturation level, but because of phenomena associated with the thawing and frost-heaved soil, the pore-water is under low or sub-atmospheric pressures in much of the soil. Relatively high pore-water pressures must occur near the surface where water flows over it, and necessarily where upward flow is evidenced by soil particles washed out from the soil (see below).

EFFECT OF FROST HEAVE ON COHESIVE STRENGTH

The distribution of void space is unusual in thawing soils. It is clear that the numerous planes of discontinuity left by ice layers and the disruption caused by small ice masses and crystals represent a very severe disturbance and an increase in void ratio, that would greatly reduce cohesion phenomena. The complex phenomenon of strength loss following frost heave is as yet little studied. It is also possible, for example, that the desiccation of clay layers between discrete ice layers caused by the suction of the developing layer produces high shearing strengths within these layers (because of the extremely low pore-water pressures), but in most cases, the effect of such strength increases must be far outweighed by the loss of strength due to discontinuities in the soil mass.

Melt-water flow.—The evidence for high permeability introduces the possibility of melt-water flow within the slope occurring with enough speed as to introduce considerable movement of soil grains. Fine sand and silt particles were also observed being carried downslope in surface water.

GENERAL CONCLUSIONS

Solifluction of this type cannot be considered analogous to soil failures due solely to increase in water content over an impermeable layer.

1. The pore-water pressures in the thawed layer are generally far lower than would occur, could the soil be saturated to the same degree in an unheaved condition.

This results partly from the drainage and void ratio characteristics of the soil in the newly thawed condition; tensions are also characteristic of unfrozen water existing in the frozen layer.

2. A considerable part of the loss of strength at thaw in frost-heaved soil is due to the reduction in the cohesion component following the separation of particles by frost heave.

3. The shear stresses are not increased proportionately to the weight of excess water present under thawing conditions, since this is counteracted by the decreased dry density of the soil.

REFERENCES

Aitchison, G. D., 1957, The strength of quasi-saturated and unsaturated soils in relation to the pressure deficiency in the pore-water: Fourth Internat. Conf. Soil Mech. Found. Eng. Proc., London, p. 135-139.

Beskow, G., 1935, Tjälbildningen och Tjällyftningen (with English summary): Sver. Geol. Unders., ser. C., no. 375, Stockholm, 242 p. Also available translated—Soil freezing and frost heaving, with special application to roads and railroads: Northwestern Univ., Evanston, Illinois, 1947.

Bjerrum, L., 1954, Theoretical and experimental investigations on the shear strength of soils: Norwegian Geotech. Inst., Publ. no. 5, Oslo, 113 p.

Det Norske Meteorologiske Institutt. Årsbok, 1949-1955 (modified average values for Fokstua).

Hardy, W. B. and Hardy, T. V., 1919, Note on static friction and on the lubricating properties of certain chemical substances: Philos. Mag., London, v. 39, no. 223, p. 32-35.

Hvorslev, M. J., 1937, Über die Festigkeitseigenschaften gestörter bindiger Böden: Ingeniørvidenskabelige Skrifter A, Nr. 45, København.

Johnson, A. W., 1952, Frost action in roads and airfields—a review of the literature: Highway Research Board, Special Report no. 1, 287 p.

Lovell, C. W., Jr., 1957, Temperature effects on phase composition and strength of partially-frozen soil: Highway Research Board, Bull. 168, Fundamental and practical concepts of soil freezing, p. 74-95.

Penner, E., 1956, Soil moisture movement during ice segregation: Building Research, Nat. Research Council, Res. Paper no. 32, Reprinted from Highway Research Board Bull. 135, p. 109-118.

Strøm, Kaare, 1952, Landskap og Istider: Norsk Turistforenings Årb., p. 48-52.

Taylor, D. W., 1948, Fundamentals of soil mechanics: John Wiley, New York, 700 p.

Terzaghi, K., 1950, Mechanism of landslides: Eng. Geol. Berkey v., Geol. Soc. America, p. 83-123.

Tsytovich, N. A., 1957, The fundamentals of frozen ground mechanics (new investigations): Fourth Internat. Conf. Soil Mech. Found. Eng. Proc., London, p. 116-119.

Tsytovich, N. A., 1947, The unfrozen water in porous strata: Izvestia of Academy of Sciences, Geol. Ser. no. 3, p. 39-48.

West, L., 1953, Determination of the dry density of compacted layers of coarse granular materials: Surveyor, London, v. 112 (3205), p. 531-532.

Williams, P. J., 1957a, Some investigations into solifluction features in Norway: Geog. Jour., v. 123, pt. 1, p. 42-58.

———, 1957b, The direct recording of solifluction movements: Am. Jour. Sci., v. 255, p. 705-714.

———, in press, The development and significance of stony earth circles.

Williams, P. J., et al., 1957, N.E.A.T.: Second report of investigations into certain solifluction and patterned ground features in Norway: Oslo, Bibliotekkontoret, Universitet, Blindern, 41 p. (cyclostyled and privately distributed).

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GEOMORPHOLOGY AND CRUSTAL MOVEMENTS OF THE ARU ISLANDS IN RELATION TO THE PLEISTOCENE DRAINAGE OF THE SAHUL SHELF

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ABSTRACT. The Aru Islands form an oblong geanticlinal upwarp of the Sahul shelf to the west-northwest of Australia. Many contradictory opinions have been expressed so far concerning their lithological structure and their recent tectonics. The author's conclusions —based on field observations and on a study of photogrammetric maps—are preceded by a discussion of the work of former investigators.

Geomorphologically the archipelago is a low plateau with slight undulations. Structural terraces are common and have been wrongly attributed to Recent uplifting by several authors. Sunken coast lines and drowned abrasion platforms indicate subsidence of the outer zones in Recent times.

The most remarkable geomorphic phenomena of the group, i.e. the channels or "sungi" between the islands, are explained as a result of a pattern of diagonal shear joints. The youthful appearance of part of these joints points to a continued warping of the Aru rise in sub-Recent or even Recent times. Neither the channels nor the submarine relief of the Sahul shelf have any connection with the Pleistocene courses of New Guinea rivers, as has often been suggested.

The Aru Islands, situated in the eastern part of Indonesia, are located on the Sahul shelf near the Australian continental margin (fig. 1). Thus to the east of the group the sea floor does not exceed 20 meters, whereas approximately 30 kilometers to the west the depth suddenly increases, and it reaches 3650 meters in the central parts of the Aru basin. The islands form part of a gentle, east-west running submarine ridge, the continuation of the so-called Oriomo axis (Carey, 1938) of eastern New Guinea and the Merauke Ridge (Van Bemmelen, 1949) of western New Guinea. Roughly speaking the Aru group has the shape of a huge ellipse, the longer axis having an azimuth of approximately 0-10° and being about twice the shorter axis. H. A. Brouwer (1925) maintains that the Banda orogene influenced the neighboring shelf and thus caused the Aru rise. This explains, according to this author, why the axis is more or less parallel to the nearest island of the Banda outer arc, Key Major. The whole 8600 sq.km of the archipelago is a low plateau, with slight local undulations. The maximum height is about 240 meters, but the bulk of the region is much lower.

The islands are built up of Neogene and Quaternary marls, soft limestones and sandstones (Gregory, 1924). The beds are subhorizontal throughout the group, and only locally a few gentle dips are observed. Ph. H. Kuenen (1933) measured gentle westward dips at the west coast between the Manumbai and Workai channels, whereas R. Tayama (1936) mentions a northwestward dip of the island of Wamar in the northwestern part of the archipelago. Thus a geanticlinal uparching is suggested. The author's observations of southwestward dips near Cape Lelar in the extreme south western part of the group agree with this, although the gentle dips measured in the row of islands to the east of the Aru "mainland" point to a somewhat more complicated development there, i.e. a slight folding.

The sediments mentioned above form only a thin cover, and the basement rocks must be present at shallow depth, as appears from the occurrence

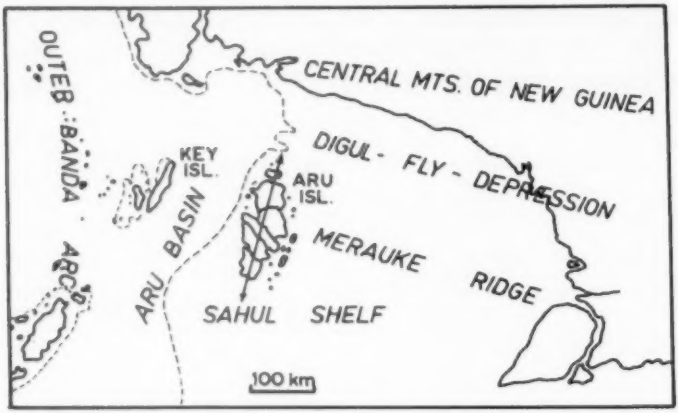


Fig. 1. Location map of the Aru islands showing the main geotectonic regions. The Australian continental margin is indicated by the 200 meters depth contour.

of coarse terrigenous minerals (quartz, feldspar, mica, etc.) (Fairbridge, 1951). Tissot van Patot (1908) observed an outcrop which he thought was probably granite, near the village of Sia in the southeast part of the group. This was not confirmed by A. Heim (van Bemmelen, 1949) nor by the present author who visited the spot at the end of 1956.

R. D. M. Verbeek (1908) considered the Aru Islands to be a raised coral reef, but this opinion is not in accordance with the present geological knowledge. Tayama, on the strength of a few coral remains found on the island of Wamar, still maintains that the archipelago is a raised plateau reef, but this evidence appears rather insufficient for such a conclusion.

The abundance of living reef has been mentioned as an argument in favor of the raised reef hypothesis. This theory being abandoned, Kuenen looked for an explanation of the lack of raised reefs in this uparched region surrounded by flourishing living reefs. He visited 60 kilometers of the west coast and found that the reef is not as abundant as usually assumed. A vast reef flat, 15-40 kilometers wide, is indicated, however, on the charts of the east coast and is also mentioned by Sperling (1936) and van Bemmelen. Kuenen therefore supposes that the islands have risen above the shelf during the Pleistocene regression or that the subsoil was unsuitable for coral growth. Fairbridge holds similar views, stating, on zoogeographical evidence, that the islands had not subsided after the Pleistocene and thus got no reef cap. Field observations of the present author on the shallow platform to the east of the Aru Islands revealed the interesting fact that marls, similar to those occurring on the islands, are predominant. Locally some scattered coral polyps are found, but shell accumulations—mostly oysters—are far more common. Thus the conditions for coral upgrowth nowadays seem to be as unfavorable as they were in the earlier stages of the uparching of the Aru Islands. The soft marls apparently are not a suitable environment for the polyps.

The flats can only be considered as an abrasion platform, and because the

seaward parts now are situated several meters below mean sea level, recent subsidence of the eastern part of the archipelago is obvious. This observation invites a study of the Recent and sub-Recent crustal movements of the region. Quite contradictory opinions have so far been expressed on this subject. Zwierzycki (1927) mentions the large estuaries reaching far inland along the river-like channels or "sungi" between the islands. From this observation and from the mangrove-covered mud coasts he infers a recent subsidence of the whole Aru group. Sperling, on the other hand, maintains that the area was subjected to a recent upheaval, with the possible exception of the northern parts. Two arguments in favor of this view are Merton's observations (1910) of raised coast lines and notches in the interior of the islands and Tissot van Patot's observations of raised coast lines and notches on southwestern Terangan. Field observations showed, however, that these phenomena are the result of minor lithological differences of the subhorizontal beds and thus do not suggest regional uplift. In areas of dipping strata, the "notches" are also tilted. The very active recession of these small cliffs is another indication that selective erosion instead of wave action is their cause.

Fairbridge suggests an eastward tilt of the archipelago because of the predominance of mangrove swamps, isolated islands, etc., in the east, and the occurrence of low cliffs along the west coast. Subsidence of the east coast is evident from the features just described, but uplift of the west coast is less obvious and is in contradiction with both the westward dip of even the youngest beds and the embayed coast line. The occurrence of cliff coasts does not necessarily indicate emergence. On the contrary, subsidence will usually result in cliff coasts, whereas coastal plains are formed along emerging coasts. The red cliffs of the southeast coast of the island of Terangan are a good example of a subsiding coast that quickly reached old age in the feebly resistant rock (pl. 1). The differences between the east and the west coast are easily accounted for by the asymmetry of the geanticlinal upwarp. This asymmetry is also reflected in the relief, the greatest heights being observed near the west coast whereas the eastern part is a low plateau region.

A study of the numerous joints occurring on the islands enabled the author to conclude that the warping of the Aru rise continued to sub-Recent or even Recent times, thus resulting in a sinking of the coastal areas and possibly in a minor rise of the interior. To explain this the drainage pattern of the archipelago needs some further consideration.

The occurrence of a number of river-like channels running across the group and dividing it into islands is beyond doubt the most remarkable geomorphic phenomenon of the Aru Islands. Numerous branch channels are also encountered. There are several theories concerning the genesis of these channels. Wallace (1857, 1869) tried to explain them as the remainders of the Pleistocene lower courses of New Guinea rivers preserved here by subsequent uparching of the Aru region, whereas elsewhere the river courses gradually disappeared during the transgression of the shelf associated with the postglacial rise in sea level. Van Hoëvell (1890) discarded this river hypothesis because of the very irregular cross and longitudinal profiles of the "sungi". The great depth of the channels—locally even deeper than the surrrounding shelf—is an-

PLATE 1

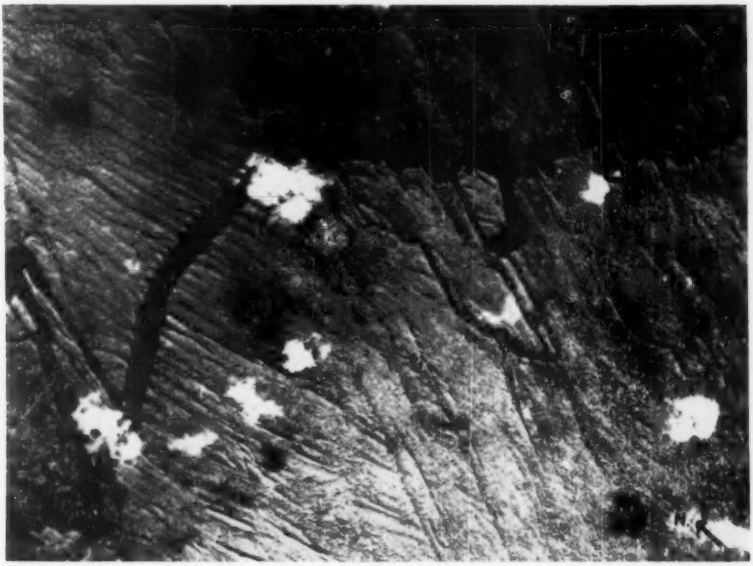
Cliffs on the old shore line of submergence in the SE part of the Aru islands (Terangan Island).

other argument. He maintains that faults resulting from the uplift of the area are the cause of the pattern of channels. Verbeek and Brouwer share this opinion.

Fairbridge, using air photographs, considers that jointing had some influence but, on the whole, he adheres to a modified river hypothesis. He maintains that the Merauke Ridge acted as a divide of the Pleistocene Sahul shelf and prevented the New Guinea rivers from flowing across the Aru region. Thus he supposes that comparatively short rivers, having their sources slightly east of the present archipelago, are the cause of the "sungi". Yet another view was expressed by Ribbe (1888), Tissot van Patot and Tayama, who suggested a submarine origin of the "sungi", i.e. as channels in the living reef, modified by the tidal currents. Sperling rejects this theory because the Aru Islands remained above sea level after the Pleistocene epoch; otherwise the faunal relationships are incomprehensible. This author, in fact, is not satisfied with any of the theories, but stresses the importance of tidal currents and further mentions depression zones as a primary cause, but without attempt at proof.

A series of Japanese photogrammetric war-time maps throws new light on the question. It reveals an abundance of criss-cross joints resulting in an angular pattern of the "sungi", narrow swamps, etc. It appears that the location and direction of the branch channels are fully imposed by these joints. The same can be ascertained for the greater portion of the main "sungi", the remaining part being uncertain due to changes and enlargements by tidal currents and

PLATE 2



Aerial view of the drainage pattern along the Maikur channel, showing strong influence of jointing. Scale 1:40,000.

solution. Beyond doubt the drainage pattern is the direct result of jointing (pl. 2).

A study of the frequency distribution of joint directions yielded results that give further insight into the crustal movements of the area. The azimuth and the length of every rectilinear drainage element was measured on the eight sheets of the 1:100,000 photogrammetric map. The azimuths were grouped together per 5°, whereas the lengths were indicated in map centimeters which equal kilometers in the field. In total, 527 directions were measured amounting to 2086 kilometers joint length. There are no photogrammetric maps of the southern extremity of the archipelago, so this area has not been considered. The data are given in table 1.

It would have been preferable to carry out the measurements on the air photographs. Unfortunately, however, the Japanese vertical air photographs were not at our disposal, whereas the available trimetrogon war photos are not suitable for our purpose due to their low quality and incomplete coverage. Measurement in the field of such a large number of joint directions and lengths would have been impossible because of the dense vegetation, the mangrove swamps bordering the coasts, etc. Thus only the possibility of map measurements remained, with the inevitable consequence that a certain number of rectilinear drainage features will be wrongly interpreted as joints and thus be included in the measurements. This is not a serious handicap, however, because their distribution will be at random and will not influence, therefore, the out-

TABLE 1

Frequency Distribution of the Joints per Map Sheet
(The lengths are indicated in map centimeters whereas the azimuths are grouped per 5 degrees).

Mapsheet	I	II	III	IV	V	VI	VII	VIII	I/VIII
Azimuth									
05	6	0	3	4	0	0	1	0	14
10	0	1	2	3	1	0	5	10	22
15	0	9	0	12	1	0	6	5	33
20	4	34	11	0	13	5	2	9	78
25	2	59	1	5	11	4	143	27	252
30	11	4	19	41	35	4	22	0	136
35	2	15	29	76	3	3	8	7	143
40	13	23	17	19	32	3	40	4	151
45	7	12	9	75	45	15	38	0	201
50	9	0	6	8	21	0	0	0	44
55	0	10	21	15	16	0	0	0	62
60	0	3	16	0	3	9	4	0	35
65	13	0	7	15	0	0	0	0	35
70	7	0	0	4	6	0	0	0	17
75	0	0	0	0	0	0	0	0	0
80	2	4	0	0	0	0	0	0	6
85	0	0	0	0	0	0	0	0	0
90	0	0	0	4	0	0	0	0	4
95	0	6	6	0	0	0	2	0	14
100	0	0	2	0	0	0	0	0	2
105	0	3	0	0	2	0	0	0	5
110	0	0	1	0	0	0	0	0	1
115	2	0	7	0	0	0	0	0	9
120	7	22	7	3	0	3	0	0	49
125	4	5	4	10	1	0	0	0	24
130	2	21	19	13	6	4	0	0	65
135	2	21	5	4	3	0	12	0	47
140	6	4	10	10	9	7	0	3	49
145	0	12	15	23	31	11	8	3	103
150	5	36	13	18	48	2	8	13	143
155	17	10	12	40	14	0	7	4	104
160	0	20	8	12	4	4	23	16	87
165	5	0	6	0	13	0	33	0	57
170	13	0	2	9	0	1	0	0	25
175	0	0	0	10	0	2	0	1	13
180	5	2	13	16	13	7	0	0	56
Total length	144	336	278	449	331	81	365	102	2086

come of the joint study. It is likely that a few pseudo-joints will be measured in every direction and thus the minima will be less pronounced. The maxima also will be broader and flatter. To eliminate the influence of meaningless local irregularities and to abstract only the general picture, the running averages were computed for every five 5°-group; this smoothing was carried out twice.

The results per map sheet are indicated by the 8 graphs of figure 2. Because the number of joints varies widely for the several map sheets, these 8 graphs were all reduced to the same superficies in order to make them more easily compared.

Finally the measurements for the whole archipelago were used to construct the strike frequency diagram of figure 3 (frequency is measured in map centi-

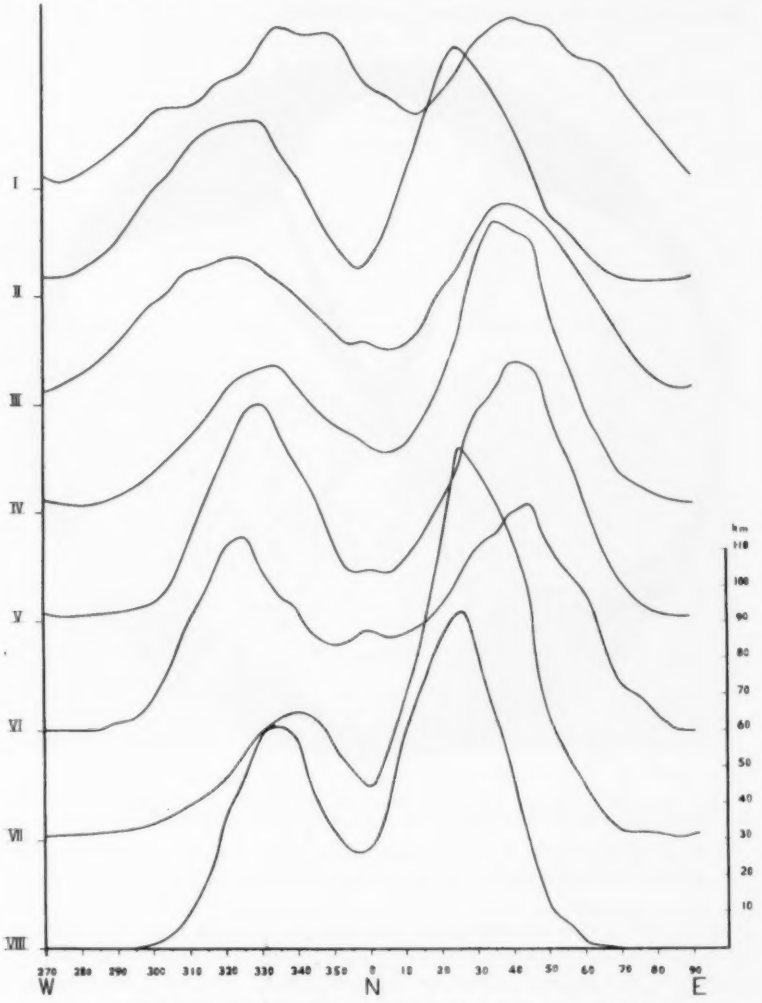


Fig. 2. Doubly smoothed graphs of the joint frequency of the Aru islands, for every map sheet. For location of the map sheets see fig. 4.

meters, not numbers of joints). The double smoothing is indicated around the diagram and pointing inward. It is evident that the maxima and minima of the direct observations are not shifted in the double smoothing, but that the whole curve is flattened out whereas secondary maxima and minima have disappeared. The number of observations was considered insufficient to allow for the construction of such a strike frequency diagram for every map sheet, but the doubly smoothed graphs of figure 2 nevertheless enable a fairly accurate determination of the predominant joint directions.

Fig. 3. Joint frequency diagram of the Aru group. The joint system is symmetric with respect to the longer axis of the archipelago, which runs approx. N-S.

As is clearly shown by the graphs, there exists a system of two diagonal joint directions making an acute angle (approx. 65°) in the north and south, and being symmetrical with respect to the axis of the islands. The northeast-southwest joints are more frequent than the other predominant direction. In between these maxima the number of joints decreases to very low values, and from this it follows that only a very limited amount of the measured drainage features has no relation to jointing and therefore should not have been included. The results thus seem to be fairly reliable.

Further consideration of the graphs reveals a number of local differences. The location of the maxima and minima is not exactly the same on all eight map sheets. The symmetry with regard to the axis of the archipelago remains, the only change being in the acute angle between the predominant directions. Especially on the northern and southern map sheets this angle becomes more acute. The minimum in between also underwent minor changes with the general tendency of being slightly shifted westward in the western part of the group and eastward in the eastern part. Sometimes an unimportant small maximum occurs in the direction of the axis. The explanation of these minor irregularities is hard to give, but certainly they result from local causes.

As a whole the phenomena observed strongly resemble a pattern of shear joints on a geanticlinal upwarp. The youthful appearance of a large number of the joints points to their rather recent formation, in view of the quick solution

PLATE 3

Bowl-shaped solution features along a minor joint in the Aru islands.

of the marls. It therefore seems clear that the warping continued until recently or is even still active. Plate 3 depicts a minor joint. As stated earlier, these tectonic movements resulted in a subsidence of the coast line and possibly in a minor uplift in the interior.

The number of joints varies widely for the individual map sheets. Even if we allow for the different percentages of the areas covered by sea, sheet 1 in the northwest portion of the group contains more than five times as many kilometers joint length as sheet 6. The northern extremity is by far the most disrupted part of the Aru group, and the southeast part is the least (fig. 4).

Doubtless these differences are caused by variations in the lime content of the sediments. There is less jointing in the sandy southeast part, especially the island of Terangan. Moreover, joints once formed, will develop less quickly because of the lower rate of solution due to the lesser lime content. The relation between the number of joints and the lime content is only a qualitative

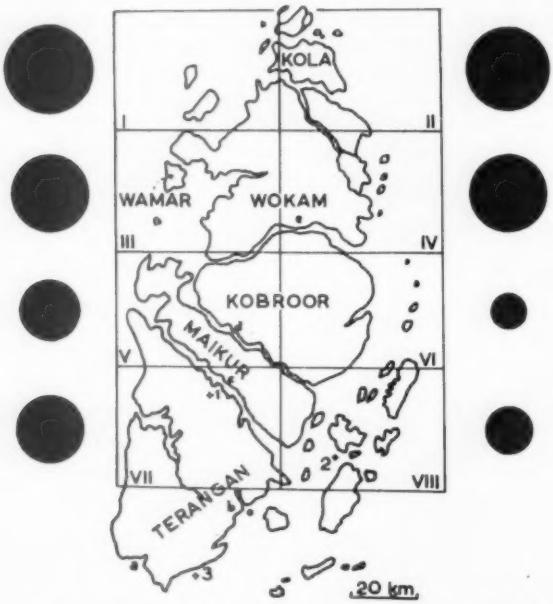


Fig. 4. Map indicating the number of joints for every map sheet. The sandy SE area is the least disrupted part of the islands. Key: a. Cape Lelar; b. Sia; c. Maikur Channel; d. Workai channel; e. Manumbai channel. 1-3: Location of the photographs.

one. In the areas of high lime content, channels will develop easiest, and because of their surroundings, solution will be most advanced and more joints will be visible there. Thus the picture induced by the differences in lime content is strongly exaggerated by subsequent solution, and the number of joints therefore bears no quantitative relation to the percentage of lime.

After this explanation of the channel pattern as imposed by a joint system, invoked by uparching, the Pleistocene drainage pattern of the Sahul shelf deserves some further consideration. Fairbridge rejects Wallace's theory and maintains that the Merauke Ridge already existed in Pleistocene times. Thus the New Guinea rivers at that time were forced to flow westward in the Digul-Fly depression. The rivers now debouching to the east of Frederik Hendrik Island took in the Pleistocene a course to the south of the Aru group.

The Digul-Fly depression is still subsiding, as indicated by the well-known drowned forests of southern New Guinea. Though this zone existed already in the Pleistocene, there is no definite proof that it formed, topographically, a

depression. This, of course, depends on the equilibrium at the time, between subsidence and sedimentation. Zoogeographic evidence points to the existence of the Merauke Ridge at the time of the postglacial transgression of the Sahul shelf, but its presence during the older Pleistocene is uncertain. Fairbridge's view, however, is supported by recent echo soundings. Though the number of soundings is not yet sufficient to allow for the construction of the Pleistocene drainage pattern, it is clear from the northwest-southeast directed sections that no former river valleys cross the Merauke ridge.

The sudden increase in depth recorded off the south coast of New Guinea shows that the recent sedimentation of the New Guinea rivers is restricted to a narrow zone along the coast and thus the Pleistocene river courses in the Digul-Fly depression will still be discernable, provided that a sufficient number of soundings is available. The edge of the continental shelf to the north of the Aru Islands appears to be rather irregular. A submarine ridge, running from east to west, occurs between two deeper belts, the latter probably being two Pleistocene river courses. The southernmost submarine valley can be traced as a valley, 100 meters deep and running north-northeast as far as approximately 25 kilometers northeast of the Aru group.

On the whole not much remains of Wallace's river theory on the peculiar "sungi" of the Aru Islands. There is no reason to assume the existence of short east-west directed Pleistocene rivers either. If rivers ever have occupied the main "sungi", it was only an incidental result of the solution along the joints, and they were not the cause of the channels. The geomorphology of the islands, especially the coastal development and the drainage pattern, is strongly influenced by young crustal movements, i.e. a continued warping of the Aru rise.

References

Brouwer, H. A., 1925, The geology of the Netherlands Indies: New York, Macmillan Co., 160 p.

Carey, S. W., 1938, The morphology of New Guinea: Australian Geographer, v. 3, p. 3-31.

Fairbridge, R. W., 1951, The Aroe Islands and the continental shelf north of Australia: Scope, Science Union Jour., Univ. Western Australia, v. 1, p. 24-29.

Gregory, J. W., and others, 1924, The Aru Islands: Geol. Mag., v. 61, p. 52-72.

Kuenen, Ph. H., 1933, Geology of Coral Reefs: Results of the Snellius expedition, v. 5, Geological Results, pt. 2.

Merton, H., 1910, Ergenbisse einer zoologischen Forschungsreise in den südöstlichen Molukken: Senckenbergischen Naturf. Gesell. Abh., v. 33, p. 1-208.

Ribbe, C., 1888, Die Aroe Inseln: Ver. f. Erdkunde zu Dresden Festschr.

Sperling, I., 1936, Beiträge zur Länderkunde von Niederländisch Neu Guinea: Frankfurter Geog. Hefte, v. 10, no. 1.

Tayama, R., 1936, Topography, geology and Coral Reef of the Aru Islands in the Eastern Dutch Indies: Inst. Geol. Palaeont. Contrib., Tohoku Imperial Univ., v. 20. (English abstr. Japanese Jour. Geol. Geog., 1939, v. 16, p. 31-32.)

Tissot van Patot, J. W., 1908, Een viertal tochten door het eiland Terangan (Aroe eilanden) in Maart en April 1907: Koninkl. Nederlandsch Aardrijksk. Gen. Tijdschr., v. 25, p. 77-93.

van Bemmelen, R. W., 1949, The geology of Indonesia: The Hague, Government Printing Office.

van Hoëvell, G. W. W. C., 1890, De Aroe-eilanden, geographisch, ethnographisch en commercieel: Indische Taal-, Land-, en Volkenkunde Tijdschr., v. 33, p. 57-102.

Verbeek, R. D. M., 1908, Molukkenverslag: Jaarb. v. h. Mijnwezen in Ned. Indië, v. 37, wetensch. gedeelte.

Verstappen, H. Th., 1959, A contribution to the geomorphology of the Moluccas: Geog. Inst. Pub., Djakarta.

Wallace, A. R., 1857, On the natural history of the Aru Islands: Ann. and Mg. Nat. History, ser. 2, v. 20, p. 473-485.

Zwierzycki, J., 1927, Toelichting bij blad XXI van de geologische overzichtskaart v. d. Ned. Ind. archipel, 1:1,000,000: Jaarb. v. h. Mijnwezen in Ned. Indië v. 56, Verh. 1, p. 308-336 (with English summary).

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THE GEOMORPHIC SIGNIFICANCE OF SOME OXFORD SOILS

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ABSTRACT. An investigation of the physical properties of soils derived from the Upper Greensand, Shotover Sand, Kimmeridge Clay and Oxford Clay formations in the Oxford region, England, reveals striking differences between them. Soil shearing resistances, measured with a Vicksburg needle-type penetrometer, can be related to a theoretical Index of Resistance derived from numerical values of soil moisture, density and range of grain sizes. Combining this Index with an estimated Coefficient of Permeability, the relative resistance of each soil type to sheet erosion can be estimated as an Index of Erodibility, which is found to bear an impressive relationship to the relief maintained by each geological formation.

NATURE OF THE INVESTIGATION

The importance of the physical properties of soils, in areas where soil-forming processes can keep pace with the erosive ones of sheetwash, creep and mass movements, has been consistently ignored by geomorphologists, except on the most superficially qualitative level. Yet, in a soil-covered region it is the stress necessary to shear off the surface soil particles which is more significant from the viewpoint of erosion than that necessary to fracture the associated bed-rock; and, except under the most intense or prolonged rain which results in the clogging of the whole depth of the soil column, it is the permeability of the soil layers ('infiltration capacity' of Horton, 1945) which controls surface run off—and therefore the intensity of surficial sheet erosion—rather than the permeability of the underlying rock. In these instances, bedrock is not to be considered a parent of the related topography, but rather a grand-parent. It is quite possible for areas of differing bedrock to present, under comparable climatic and vegetative conditions, landforms of similar geometry, simply because both rock types have weathered to produce soils having similar physical properties. As G. K. Gilbert sagaciously noted more than eighty years ago:

> We have seen that vegetation favors the disintegration of rocks and retards the transportation of the disintegrated material. Where vegetation is profuse there is always an excess of material awaiting transportation, and the limit to the rate of erosion comes to be merely the limit to the rate of transportation. And since the diversities of rock texture, such as hardness and softness, affect only the rate of disintegration . . . . and not the rate of transportation, these diversities do not affect the rate of erosion in regions of profuse vegetation, and do not produce corresponding diversities of form. (Gilbert, 1877, p. 119)

With this thesis in mind, the physical properties of the soils derived from four geological formations in the Oxford region (the Upper Greensand, Shotover Sand, Kimmeridge Clay and Oxford Clay) were investigated in order to:

1. Determine the distinct physical properties of the four soil types associated with geological formations exhibiting a variety of relief forms.
2. Obtain quantitative measurement of the shearing resistances of the soils, and to examine the characteristics of their statistical distribution.
3. Examine the relationships between the measured soil shearing resistances and the physical soil properties.
4. Attempt some general rational explanation of relative local relief in the light of the resistive properties presented by the associated soils to sheet erosion.

Three sample areas were chosen on each of the four formations (fig. 1), such that the ground slope at each is not significantly different from zero—for

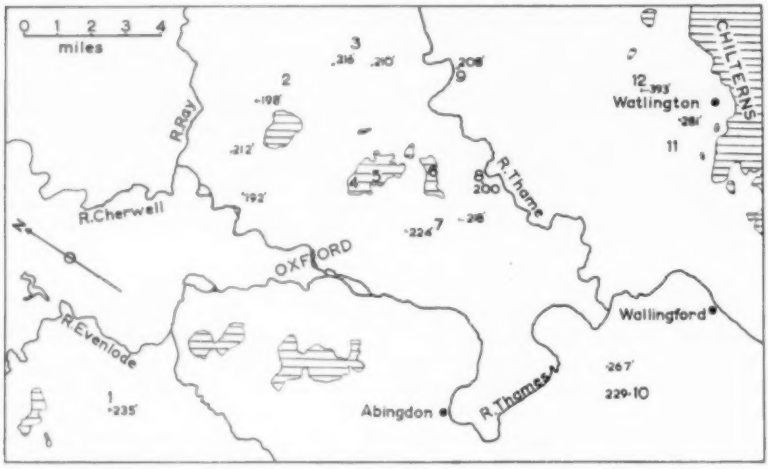


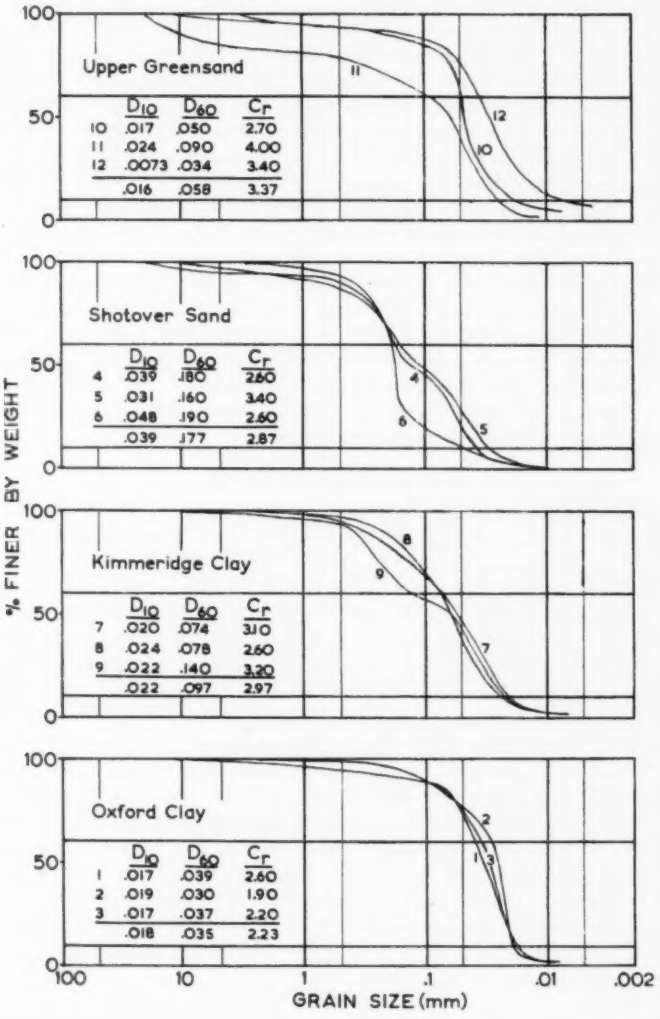
Fig. 1. Location of the twelve localities. Land over 400 feet shaded.

differing rates of creep on different slope angles may well cause soil density and shearing resistance to vary between otherwise similar soils—and such that all the localities were composed of permanent or long-ley pasture, in an attempt to reduce as far as possible the differential binding effects of vegetation of differing type or age. Thus, all readings were made with respect to closely-grassed level surfaces, in order to eliminate gross resistive differences caused by differential vegetative soil binding. At each locality fifty 6-inch deep penetration readings were made to give a quantitative measure of the soil shearing resistance; a disturbed sample of the top 6 inches of soil was taken for grain size and moisture analyses; and undisturbed core samples were taken as a measure of soil density and so that, in one instance (that of the Shotover Sand), permeability tests could be conducted under field conditions. Field work was carried on intermittently during the months of May, June and July 1958 under abnormally damp conditions; the Thornthwaite Monthly Precipitation-Evaporation Ratios $\left(\frac{P}{E}\right)$ for Oxford being .403, .587 and .403, respectively (Thornthwaite, 1931). Readings on different geological formations were roughly alternated in order to obtain the widest range of soil moisture conditions for each during the three month period.

GRAIN SIZE CHARACTERISTICS

Each of the twelve disturbed soil samples was examined in respect of its grain size characteristics, according to the principles developed by Burmister (1951). The distributions of grain sizes for the sieve fractions covering sand

sizes and coarser were determined by wet sieving, whereas the grain size proportions for the silt and clay sizes ($< 1/16$ mm) were obtained by using the pipette method of dispersed particle settling under Stokes' Law, explained by Krumbein and Pettijohn (1938). Figure 2 shows these grain size distributions plotted conventionally as cumulative percentages finer by weight.



	D_{10}	D_{60}	C_r
10	.017	.050	2.70
11	.024	.090	4.00
12	.0073	.034	3.40
	.016	.058	3.37

	D_{10}	D_{60}	C_r
4	.039	.180	2.60
5	.031	.160	3.40
6	.048	.190	2.60
	.039	.177	2.87

	D_{10}	D_{60}	C_r
7	.020	.074	3.10
8	.024	.078	2.60
9	.022	.140	3.20
	.022	.097	2.97

	D_{10}	D_{60}	C_r
1	.017	.039	2.60
2	.019	.030	1.90
3	.017	.037	2.20
	.018	.035	2.23

Fig. 2. Grain size distribution curves.

Burmister (1951, p. 21) has shown that the grain size characteristics of a soil can be expressed completely by the use of three parameters, two of them quantitative. These are:

1. The 'fineness' of the soil—expressed by "Hazen's Effective Size" (D_{10}), the grain size in millimeters for which 10 percent of the material is finer, by weight. The proportion of the very fine material may be an extremely important control over other physical properties of the soil, particularly over shearing resistance and permeability. Values of D_{10} together with those of D_{60}, are given in figure 2.
2. The range of grain sizes—analogous to the statistical "standard deviation". Soil engineers conventionally express this range in terms of the number of sieve fractions (where coarse, medium and fine fractions each count as one) spanned by the mean slope line of the cumulative grain size curve. This range of sizes (C_r) expresses the degree of sorting and is therefore inversely related to the permeability and directly related to the angle of internal friction (Van Burkalow, 1945) and, consequently, to the shearing resistance.
3. Type of grain size distribution curve—analogous to the symmetry or statistical "skewness" of the size distribution. The most common type, here represented by the Shotover, Kimmeridge and Oxford soils, is the 'S' curve which approximates to the symmetrical "normal" distribution. The 'E' type is typically left-skewed with an excessive admixture of coarse fractions; this is exhibited to varying degrees by the three Upper Greensand soil samples.

SOIL MOISTURE AND DENSITY

Before the grain size analyses were made, the disturbed soil samples were weighed, dried in an oven at 80°F, reweighed, and the moisture content calculated as a percentage of the dry weight. The drying time naturally varied with the clay content, but it was generally not less than 24 hours—except for the well-drained Shotover Sand. Such drying was, of course, not completely effective for the Oxford Clay soil with its high clay and organic content, but the calculated moisture contents do provide at least a reasonably satisfactory quantitative measure of the relative humidity of the soils under the conditions of field penetration measurement. The undisturbed soil samples, obtained by driving thin metal tubes (to minimize soil compression in the tube) of 2-inch diameter into the soil, were used to determine, after the contents had been dried and weighed, the dry weight density of the soil samples in gm./cu.inch.

These two soil parameters of moisture and density are closely connected, particularly for clay-rich soils, for every increase in moisture content will naturally cause a decrease in the dry weight density. This explains the low densities shown in table 1 for the highly saturated Oxford Clay soils, and underlines the necessity for both measures to be used where a composite parameter of soil behavior is required.

SHEARING RESISTANCE

Soil shearing resistance was measured in terms of the resistance presented to the direct penetration of a needle point having a 1/10 sq. inch area, with a

conical 60° point angle, manually driven into the top 6 inches of soil under a steadily increasing pressure. The instrument employed was a Vicksburg Penetrometer (fig. 3), described by Evans (1951). At each of the twelve localities

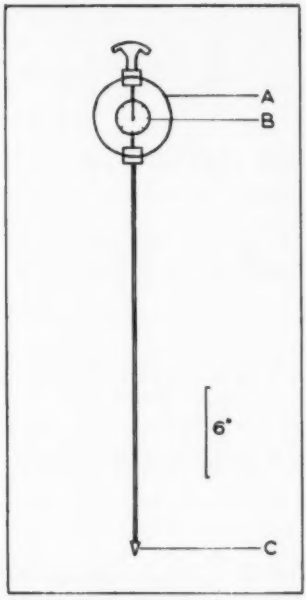


Fig. 3. The Vicksburg Soil Penetrometer.
A. Proving ring (S.A.E. 1066 steel)
B. Ames gauge.
C. Needle point or cone (1 sq. in. or 1/10 sq. in.; 60° point angle).

the penetrometer was driven fifty times into the soil surface layers by one operator, while the other read the maximum compression of the proving ring (in 1/1000s inch) as measured by the Ames Gauge. As the instrument used was not new, the proving ring was calibrated before and after the period of field measurements by compressing it under known weights, such that the final mean resistance values in 1/1000s inch could be translated into lb./sq.inch.

Histograms showing the distribution of penetration values for each locality are shown in figure 4. These indicate that for fine-grained soils of low shearing resistance the distributions tend to be normal, whereas with increasing resistances greater variation is found between differing soils of the same type, and greater standard deviations (S) occur within the readings for each individual locality. Where a coarse fraction is involved, as with the Upper Greensand, the frequency distributions tend to become right-skewed because of the occasional high readings occurring when non-compressible stones are approached by the needle. Readings when the needle actually encountered stones were naturally discarded. Statistical analyses of variance (Walker and Lev, 1953) were made between the three penetration distributions for each soil

TABLE 1
Table of All Numerical Values

Location		Mean Measured Shearing Resistance		Moisture (% Dry weight)	Density (GM/cu.in.)	D_{10} (M.M.)	D_{60} (M.M.)	C_r	Curve Type	Resistance Index (I_r)	Coeff. of Permeability (CM/Sec.) (K)	Erodibility Index (I_e)
		1/1000's in.	lb/sq.in. (R)									
Upper Greensand	10	83.02	991	21.3	18.0	.017	.050	2.7	E			
	11	67.78	814	19.5	23.0	.024	.090	4.0	E			
	12	47.88	580	20.1	20.4	.0073	.034	3.4	E(ES)			
	Mean	66.23	795	20.3	20.5	.016	.058	3.37	E	3.40	.00084	1.5
Shotover Sand	4	47.26	573	20.9	19.6	.039	.180	2.6	S			
	5	62.06	747	22.9	25.8	.031	.160	3.4	S(SE)			
	6	61.08	735	16.5	24.9	.048	.190	2.6	S(SE)			
	Mean	56.80	685	20.1	23.4	.039	.177	2.87	S	3.34	.0043	.34
Kimmeridge Clay	7	30.12	372	41.8	18.1	.020	.074	3.1	S			
	8	39.62	484	22.9	21.9	.024	.078	2.6	S			
	9	34.34	423	24.9	23.3	.022	.140	3.2	S			
	Mean	34.69	426	29.9	21.1	.022	.097	2.97	S	2.10	.0013	1.8
Oxford Clay	1	30.12	372	36.7	19.6	.017	.039	2.6	S*			
	2	24.64	309	56.0	19.0	.019	.030	1.9	S			
	3	28.24	352	41.5	18.4	.017	.037	2.2	S			
	Mean	27.67	344	44.7	19.0	.018	.035	2.23	S	.95	.00080	3.6

* Stream alluvium intermixed.

group, and in each of the four instances it was determined that the mean penetration values for each of the three soil samples of a single geological soil group differed significantly ($F > F_{.95}$). In figure 4 the penetration histograms have been arranged so that the even more significant differences between the shearing resistances of each of the soil groups is graphically apparent, the shearing resistances forming a progression from the lowest values recorded on the Oxford Clay to the highest values of the Upper Greensand. The mean penetration values have been translated into terms of lb./sq.inch in table 1, which summarizes all the numerical soil data.

FACTORS CONTROLLING SHEARING RESISTANCE

Theoretically, a number of factors can be isolated as controlling the shearing resistance of soils, and in this respect the experiments of Van Burkalow (1945) on the related angle of repose (angle of internal friction) of fragmented material are instructive. It is apparent, however, that not all soils are equally sensitive to each control and that the clay soils and granular soils are particularly distinct in this respect. Figure 5 demonstrates how the measured shearing resistances of the clay-rich Oxford, Kimmeridge and Upper Greensand soils are closely related to the field moisture percentages and how the shearing resistance of a clay soil may be drastically increased at low moisture percentages by even small decreases of moisture content. This relationship is not so apparent for a granular soil like the Shotover Sand, the shearing resistance of which is shown in the same figure to be much more sensitive to soil density variations. As has been pointed out, moisture content and dry weight density are inter-allied for clay soils, giving moisture content a multiple significance in these instances.

It is possible, partly from the work of Van Burkalow, to list the soil factors which theoretically should compositely affect soil shearing resistance, and whether these relationships are mathematically direct or inverse. They are:

1. Grain size (represented by D_{10})—inverse relationship.
2. Range of grain sizes (C_r)—direct relationship.
3. Grain roundness—inverse relationship.
4. Grain density—direct relationship.
5. Surface roughness of grains—direct relationship.
6. Particle cohesion—direct relationship. Because the cohesive clay particles have a small size, and because cohesion depends on the moisture content, this factor may be considered to enter as a secondary effect of other factors.
7. Soil density—direct relationship.
8. Soil moisture content—inverse relationship.
9. Amount of vegetative binding—direct relationship.

Of these factors, numbers 6 and 9 may be disregarded for the reasons given, and numbers 3, 4 and 5 may be eliminated because these differences between the soil particles of the different soil groups are probably very small. Thus one is left with the four quantitative factors of D_{10}, C_r, moisture content and soil density which might reasonably be expected to exert the predominant composite control over the soil shearing resistance.

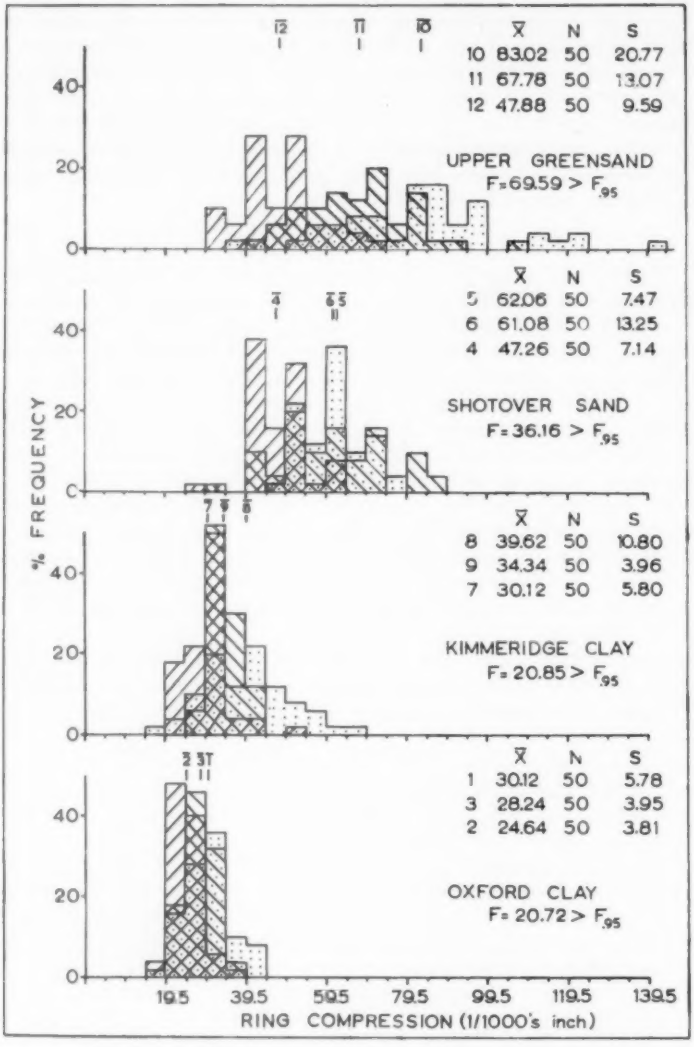


Fig. 4. Histograms of soil penetration readings.

The relationships between the mean values of these four factors and the mean measured shearing resistance for each of the soil groups are depicted in figure 6. Statistical tests of significance of trend of the regressions (Chenoweth, 1952), which were calculated by the method of least squares, indicate

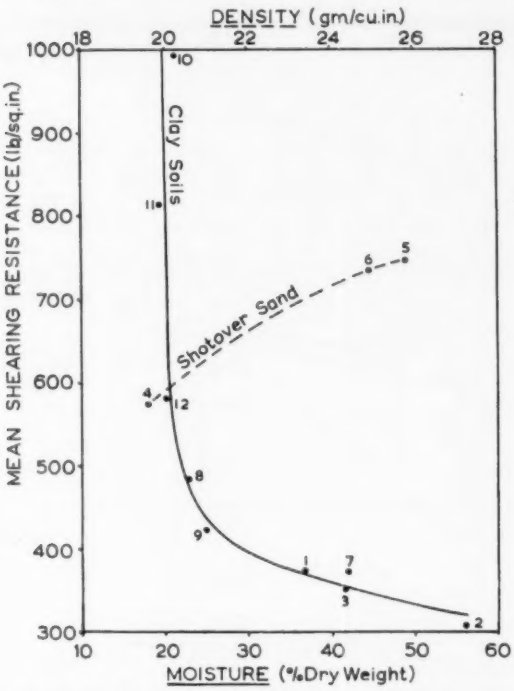


Fig. 5. Relationships between clay soil moisture content and mean shearing resistance, and between Shotover Sand density and mean shearing resistance.

that in all four instances there is a Probability (P) greater than the conventional chosen critical level of significance of .05, indicating that there is no reason to assume, under the guidance of conservative mathematical statistical methods, that any of the regression coefficients is significantly different from zero. This implies that there is no demonstrated relationship between either D_{10}, C_r, moisture content or soil density, on the one hand, and measured soil shearing resistance, on the other. This cautious reaction of the statistical machinery, however, is undoubtedly due in part to the small number of points from which each of the best-fit regressions were calculated, and it is instructive to examine the graphically presented trends further. The slope of the regression lines indicating that shearing resistance is directly related to C_r and density, and that is is inversely related to moisture content, are in accord with theoretical expectations. The low value of the regression coefficient (i.e. the slope) of the D_{10} regression, together with its weak trend being opposed to theoretical expectations, might lead one to suppose that this measure of grain size has small control over the mean shearing resistance of the four soil groups in question, and that it may be eliminated from future theoretical considerations with fair confidence. This is supported by the conclusion of Van Burka-

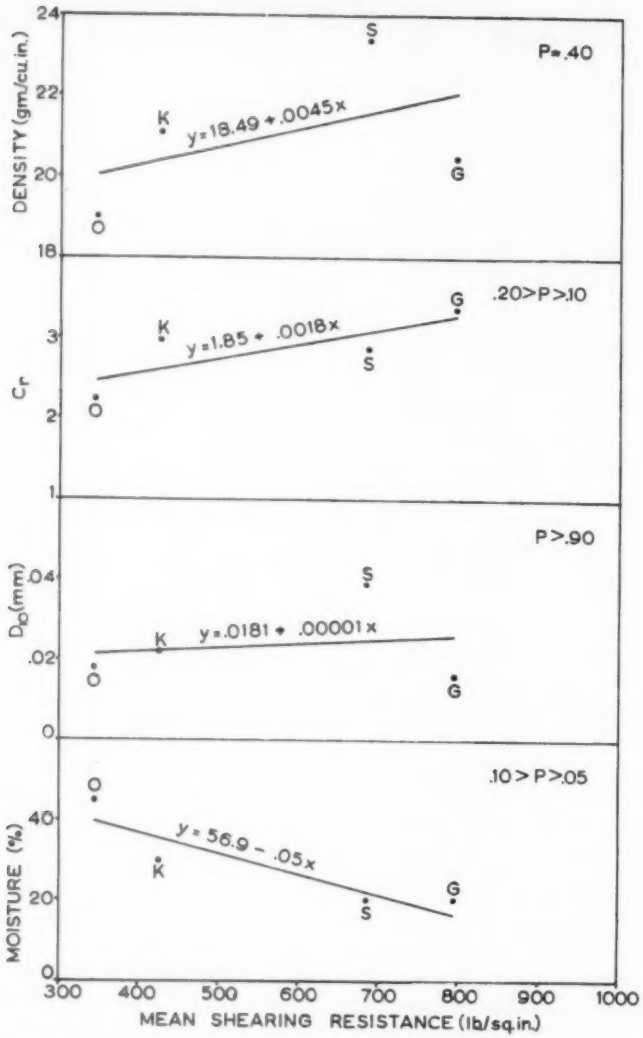


Fig. 6. Relationships between mean shearing resistance and the soil characteristics of density, C_r, D_{10} and moisture content.

low (1945, p. 703) regarding the relationship between particle size and angle of repose (angle of internal friction—directly related to shearing resistance), that: "The influence of size appears to be weak and is evident only in perfectly sorted material."

The three remaining soil parameters may be combined into a theoretical Index of Resistance (I_r) for each soil group, in the following manner:

$$I_r = \frac{\text{Soil density x } C_r}{\text{Soil moisture}}$$

The numerical values of this factor are given in table 1, and the striking relationship between them and the actual mean measured shearing resistances is shown graphically in figure 7. This relationship reassures one that the three controlling factors, which have been arrived at by practical isolation and theoretical elimination, are truly those which, for these soils at least, do control to a large extent the differential resistances which are actually presented by the four soil groups in the field.

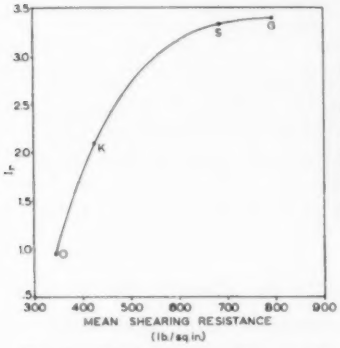


Fig. 7. Graph showing the correlation between the mean measured shearing resistance and the theoretical Index of Resistance (I_r).

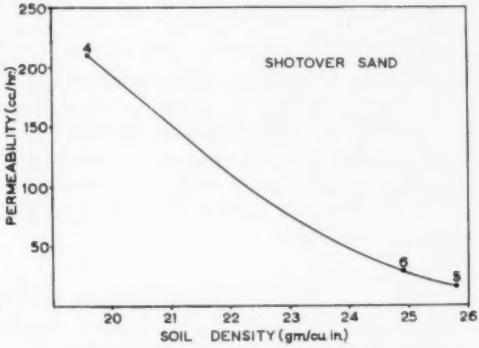


Fig. 8. Relationship between density and measured permeability for the Shotover Sand.

SOIL PERMEABILITY

Permeability is the most important control over amount of run-off at the surface under given climatic and vegetative conditions, and is one which depends largely upon measurable soil characteristics—chiefly grain size, range of grain sizes and soil density.

Equipment was not available for laboratory measurement of soil permeability, particularly of the more impermeable clay soils, but a rather crude laboratory test was made to determine the relative permeabilities of the three samples from the most permeable soil group (the Shotover Sand). 6-inch undisturbed samples were obtained from each of the three Shotover Sand localities, by driving 2-inch diameter metal tubes into the soil for half of their 1-foot length. In the laboratory the remaining 6 inches of the tubes were kept full of water (after the enclosed samples had been completely saturated) and the total quantity of water (in cc) seeping through the vertical soil columns was measured after 4 hours 10 minutes. The mean rate per hour was then calculated for each sample, and these show a striking relationship to the soil density (fig. 8), as might be expected. It was impossible to treat the less permeable

remaining soils in this manner, in the absence of equipment to produce a large head of pressure.

It is possible, however, to estimate the relative permeabilities of the four mean soil samples (averaged for each geological formation) with the aid of a graph prepared by Burmister (1951, fig. 13.4, p. 119), wherein the Coefficient of Permeability (K; measured in cm./sec.) is plotted against D_{10} and C_r for a wide range of soils, assuming a 40 percent relative density. These values are shown in table 1, and may be considered as rough relative values of the mean permeability of each soil group, despite the fact that their field densities are much greater than 40 percent (probably nearer 70 percent) and that the Upper Greensand soil is of the 'E' type size distribution, as distinct from the more normally distributed remaining three soil groups. It would seem, indeed, that the estimated permeability of the Upper Greensand soil is relatively too low.

RELATIVE ERODIBILITY OF THE FOUR SOIL GROUPS

Surface erosion by sheet run-off, under uniform climatic and vegetative conditions, is grossly dependent inversely upon the resistance presented by the soil particles to their removal by shearing, and directly upon the amount of surface run-off (i.e. inversely upon the soil permeability). It is possible, therefore, to develop a relative Index of Erodibility (I_e) for each of the four soil groups, such that:

$$I_e = \frac{1}{\text{Measured mean shearing resistance x Permeability (P)}}.$$

Values for this Index, given in table 1, for the Shotover Sand, Upper Greensand, Kimmeridge Clay and Oxford Clay are .34, 1.5, 1.8 and 3.6, respectively. This suggests that the Shotover Sand is least liable to sheet erosion and that the Oxford Clay is most susceptible to it. It seems no accident that the general relief on each of the associated geological formations falls into this order as regards its magnitude, with the Shotover Sand forming the most pronounced relief features and the Upper Greensand a more subdued ridge; whereas the Oxford Clay is associated with the major lowland of the region. These quantitative values of erodibility are, of course, only relatively indicative of general tendencies to reduction by sheetwash, for actually both the Oxford and Kimmeridge Clays are associated with low-lying areas little different in mean elevation. The approximate nature of the erodibility estimate is undoubtedly due in part to the obvious fact that sheetwash erosion is not the only mechanism of reducing the local interfluves—creep also being very important. The relative correspondence of the calculated values of erodibility with the observed relief features, however, gives reassurance that the theoretical reasoning prompting the correlation is not entirely divorced from reality, and that the geometrical features of landscape bear a much closer and more rational relationship to the physical properties of soils than has been supposed hitherto.

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References

Burmister, D. M., 1951, Soil mechanics, v. 1: New York, Columbia University, 155 p.

Chenoweth, P. A., 1952, Statistical methods applied to Trentonian stratigraphy in New York: Geol. Soc. America Bull., v. 63, p. 521-560.

Evans, I., 1951, A method for the rapid evaluation of soil strength at forward airfields: Army Oper. Res. Group., Rept. no. 21/51, 10 p.

Gilbert, G. K., 1877, Report on the geology of the Henry Mountains: Washington, 170 p.

Horton, R. E., 1945, Erosional development of streams and their drainage basins: hydrophysical approach to quantitative morphology: Geol. Soc. America Bull., v. 56, p. 275-370.

Krumbein, W. C., and Pettijohn, F. J., 1938, Manual of sedimentary petrography: New York, Appleton-Century-Crafts, Inc., 549 p.

Thornthwaite, C. W., 1931, The climates of North America according to a new classification: Geog. Rev., v. 21, p. 633-655.

Van Burkalow, A., 1945, Angle of repose and angle of sliding friction: Geol. Soc. America Bull., v. 56, p. 669-707.

Walker, H. H., and Lev, J., 1953, Statistical inference: New York, Henry Holt and Co.

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RELATIONS AMONG SODIUM BORATE HYDRATES AT THE KRAMER DEPOSIT, BORON, CALIFORNIA*

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ABSTRACT. The present relations among the hydrated sodium borate minerals at the Kramer deposit at Boron, California, when considered in the light of laboratory results on vapor pressure measurements and solubility polytherms of the $Na_2B_4O_7$—H_2O system, and of the crystal structure of borax and the colemanite series of minerals, show that the ore body was originally deposited as a chemical precipitate of borax from a saline lake at 25°-35°C and that conversion of borax to kernite took place on burial to a depth of the order of 2500 ft. and in the temperature range 53°-63°C. Crystal chemical considerations provide an adequate explanation of some of the apparently anomalous and puzzling features of the experimental physical chemical results previously found for the $Na_2B_4O_7$—H_2O system, and indicate that the formulas for borax, tincalconite, and kernite are correctly written as $Na_2B_4O_5(OH)_4.8H_2O$, $Na_2B_4O_5(OH)_4.3H_2O$, and $Na_2B_4O_6(OH)_2.3H_2O$, respectively.

INTRODUCTION

During a visit to the Kramer borate deposit near Boron, California, in April, 1957, we were shown the relations among the various hydrates of sodium tetraborate in the underground workings. Because these sodium borates are present, differing in chemical composition only in their degree of hydration, we decided to investigate the possibility that conditions of deposition and burial might be reflected by present mineralogy. The three hydrates, borax, tincalconite, and kernite, cannot coexist at equilibrium in a given environment, so it seemed likely that one or more represented "fossil" conditions.

The Kramer deposit has been described in detail previously (Gale, 1946; Schaller, 1930); here it is necessary only to recapitulate the gross aspects of the geology to provide a setting for the occurrence of the various hydrated minerals. The borate deposits are a tabular mass some 200 feet thick that dips gently beneath the nearly level floor of a broad alluvial valley, so that the overburden ranges up to several hundred feet in the area mined by underground methods. The deposits are evenly bedded, and are interlayered with thin seams of shale and carbonate muds. When the deposits are first penetrated only two sodium borate minerals are found: borax, $Na_2O.2B_2O_3.10H_2O$, and kernite, $Na_2O.2B_2O_3.4H_2O$. The occurrence of borax is reminiscent of that of halite or gypsum, in that the layers are made up of interlocking crystalline grains of borax, and show characteristics consistent with a recrystallized chemical precipitate. The grains contain small and variable amounts of impurities; grain boundaries do not cross-cut layering; and minute seams and stringers of shale occur within major beds with an attitude conformable with the major layering. Kernite, on the other hand, occurs in large part as irregular masses of large clear crystals. These masses cross-cut bedding, and in some places the mass is surrounded by a thin shaly selvage. In general, kernite occurs in greater abundance in the lower parts of the workings.

After a drift is developed, tincalconite ($Na_2O.2B_2O_3.5H_2O$) begins to form fine-grained crusts on the walls as an alteration product of borax, but

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apparently not after kernite; this fact has been noted by Muessig and Allen (1957). The change of borax to tincalconite takes place if ventilation is maintained, but in sealed workings tincalconite is reconverted to borax (oral communication, F. M. Smith, U. S. Borax and Chemical Corp.).

From these observations a number of deductions can be made that can be checked against the chemical and crystal structure relations of borax, kernite, and tincalconite.

The mode of occurrence indicates that the entire deposit was originally composed of borax. Furthermore, the even bedding and considerable lateral extent are consistent with deposition as a primary precipitate from a saline lake at temperatures probably in the range 25°-35°C. One point to be decided is whether borax is the expected phase under such conditions.

Kernite, from its cross-cutting relations, is secondary after borax. Its increasing abundance with depth suggests that dehydration of borax occurred as a result of temperature and pressure increase during burial. However, if the reaction from borax to kernite is a rapid one, kernite should be the only phase below some isogradic surface. The sporadic occurrence of kernite, plus the implication that large crystals require slow growth and difficult nucleation, are evidence for a slow approach to equilibrium. The next point to be considered is what the conditions are under which borax changes to kernite, and what is known or can be deduced concerning the reaction rate for this transition.

The apparent direct dehydration of borax, $Na_2O.2B_2O_3.10H_2O$, to kernite, $Na_2O.2B_2O_3.4H_2O$, under natural conditions as judged by the absence of tincalconite, $Na_2O.2B_2O_3.5H_2O$, when the deposits are just opened, is in striking contrast to the rapid dehydration of borax to tincalconite when exposed to Mojave Desert air by mining. These relations apparently indicate that tincalconite is the first stable dehydration product at low temperature and pressure, and that kernite is stable at some higher temperature and pressure. These observations pose the problem of finding the dehydration sequence of borax as a function of temperature, total pressure, and vapor pressure of water.

Because both dehydration of borax to tincalconite and rehydration of tincalconite to borax have been observed in the mine, it can be predicted that the reaction is easily reversible, and that equilibrium conditions for this transition are close to those observed in the working mines. The dry desert air where admitted to the mine dehydrates borax rapidly, raising the vapor pressure of water in the workings. Because so much borax is exposed it is likely that the humidity in the mine may approach the equilibrium value. When workings are sealed, the air supply is cut off, but moisture can seep in, so that the humidity rises, and tincalconite is rehydrated to borax. What, then, is the equilibrium vapor pressure for the transition, and why is the reaction easily reversible?

PHYSICAL CHEMISTRY OF THE $Na_2B_4O_7$—H_2O SYSTEM

Relations in low temperature–low pressure environments.—The basic data on the solubility and vapor pressure relations among borax, tincalconite, and kernite have been obtained by Blasdale and Slansky (1939) and by Menzel and co-workers; the work of the latter group is summarized in Menzel

and Schulz (1940), and Menzel (1935). A brief survey of the work on the $Na_2B_4O_7$—H_2O system is also given in Kemp (1956).

Menzel has found that the dehydration of borax to tincalconite may follow one or other of two different courses. Freshly crystallized borax previously heated to about 50°C and cooled will change reversibly into crystalline tincalconite, with a vapor pressure of 10 mm at 20°C. Unheated freshly crystallized borax will decompose to an amorphous compound of composition $Na_2O.2B_2O_3.2H_2O$, with vapor pressure 1.6 mm at 20°C. This latter process is not reversible. However, all observations made on the relations among the sodium borates at the Kramer deposit indicate that it is the reversible decomposition with which we are primarily concerned. The reversible vapor pressure-temperature curve for the decomposition of borax to tincalconite in the temperature range 15°C to 25°C is given in figure 1a and the corresponding relative humidity-temperature curve in figure 1b. To these curves we have added vapor pressure-temperature and humidity-temperature curves for a solution saturated with borax, and for one saturated with sodium chloride.

Equilibrium among borax, kernite, and water vapor, in the low temperature-low pressure range, apparently has never been achieved in the laboratory, and hence no experimental p-t curve for this reaction is available. Menzel and

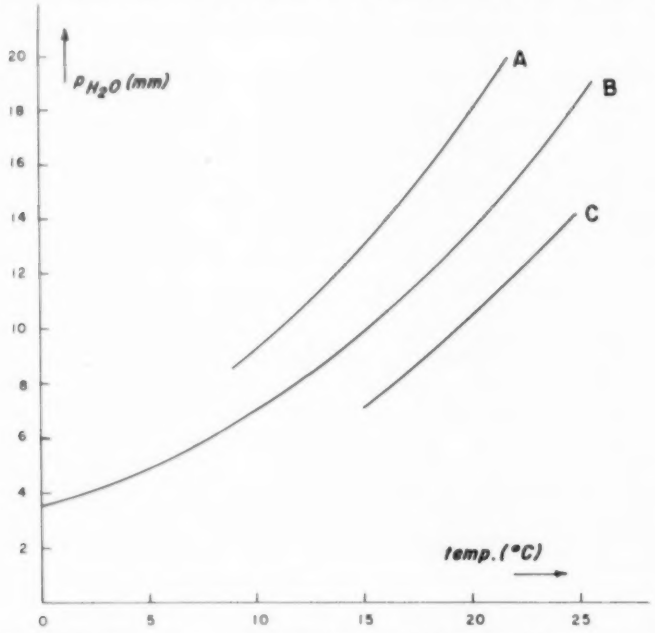

Fig. 1a. Vapor pressure-temperature plots for: (A) saturated borax solution, (B) saturated sodium chloride solution, (C) the reversible equilibrium borax $\rightleftarrows$ tincalconite. Within experimental error the p-t plot for saturated borax solution is the same as that of pure water. Data for (A) and (C) from Menzel (1935), data for (B) from *International Critical Tables*, 1933, vol. III p. 369.

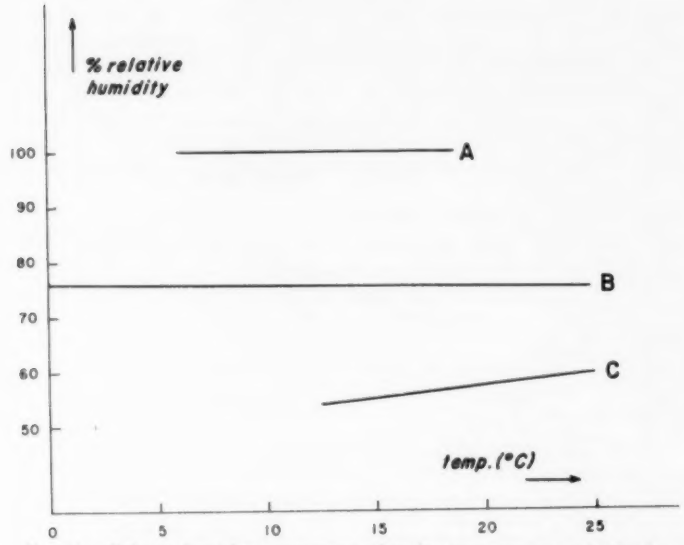

Fig. 1b. Relative humidity-temperature plots for same systems as in fig. 1a.

co-workers have made valiant efforts to calculate such a curve by indirect methods involving the use of the Nernst approximation formula which relates the heat of vaporization, the vapor pressure, the temperature, and an empirical constant (Taylor, 1925). In view of the approximations involved in the use of the Nernst equation, it seems best not to accept these p-t results at face value but rather to use them in a qualitative way. Menzel's calculations show that in the low temperature-low pressure range the p-t curve for the borax-tincalconite equilibrium would yield slightly lower vapor pressures than that for the borax-kernite equilibrium. Here, we consider it safe only to accept the fact that the two p-t curves lie near one another.

From figure 1, and the above discussion, it can be seen that on a thermodynamic basis the first dehydration product of borax could be either tincalconite or kernite. However, all evidence shows that tincalconite does grow in preference to kernite, regardless of which is the thermodynamically more stable. It can also be seen from figure 1 that the equilibrium vapor pressure for the dehydration of borax to either lower hydrate is well below that achievable in solutions saturated with borax or with sodium chloride.

Relations at higher temperatures.—If the formation of the deposits is reconstructed as deposition of borax from a solution at earth surface conditions, followed by burial and subsequent local dehydration of borax to kernite, temperature, pressure, and composition effects must be considered in deducing the probable minimum depth of burial at which kernite could form. The borax crystals at the time of deposition were bathed in a brine of unknown composition, except that it must have been saturated with respect to borax and undersaturated in other salines. Borax has a large positive coefficient of solubility

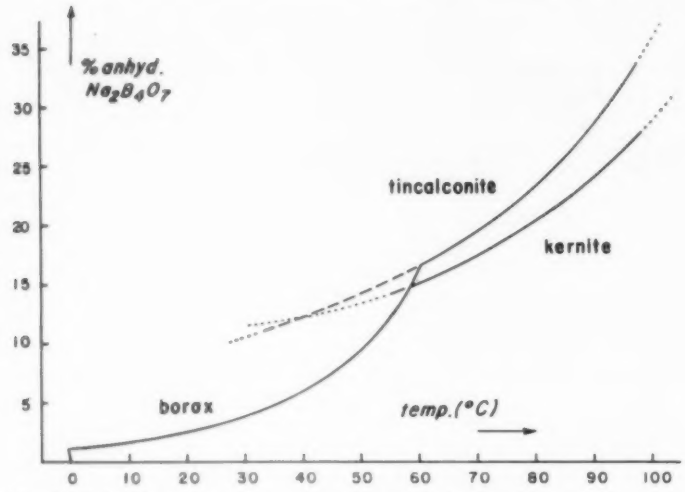

Fig. 2. Solubility-temperature curves at one atmosphere total pressure for the system $Na_2B_4O_7$—H_2O, after Menzel and Schulz (1940). The dashed portions of the curves indicate metastable lines for which observations were made, and the dotted portions are extrapolations.

(fig. 2), so that the effect of burial would be expected to be solution and recrystallization of the borax, with expression of excess brine, and the development of the dense coarsely crystalline interlocking texture now observed. With burial much of the brine was squeezed out. When kernite began to form water was released, which would tend to dilute and to express any small amount of residual interstitial brines so that as a good first approximation it can be assumed that the transition took place in a solution saturated with respect to borax and kernite but with a low content of other salines. Complications arising from the presence of partially diluted brine can be assessed separately.

The solubility-temperature curves for the system $Na_2B_4O_7$—H_2O have been given by Blasdale and Slansky (1939), and Menzel and Schulz (1940). The results of the latter authors are reproduced in figure 2. At one atmosphere total pressure borax changes to kernite at equilibrium at 58.5°C. The intersection of the borax solubility curve with that of tincalconite yields a metastable transition point at 60.8°C, and the metastable continuation of the kernite and tincalconite curves shows a metastable transition point at about 39°C. On the basis of the observations described in the introduction to the present paper there is no evidence that these two metastable transitions were in any way involved in the formation of the Kramer ore body.

At 1 atmosphere total pressure the transition temperature of borax to kernite, in a solution saturated with respect to the two borates, is 58.5°C. Other added salts in the interstitial brine would lower the transition temperature, (see, MacDonald, 1953); this effect on the transition temperature is treated separately in a following section. However, as suggested before, although the temperature of first transition to kernite might be lowered by other

salts, the low porosity in conjunction with the effect of the water produced by the reaction would soon cause the system to approach one of borax-kernite saturated solution. It may be noted that the Kramer deposit is strikingly free of salty sediments such as are found at the Tincalayu borax mine in the province of Salta, Argentina (Muessig and Allen, 1957). To a good first approximation, then, it is considered that at 1 atmosphere total pressure the minimum temperature of conversion of borax to kernite is 55°-58°C.

THE EFFECT OF UNIFORM PRESSURE

Both temperature and pressure increase with depth of burial so that the effect of uniform pressure on the transition temperature of the borax-kernite conversion must be evaluated. For the reaction:

$$\text{Borax} = \text{Kernite} + \text{Solution}$$

data are not available for the partial molal volume of borax at temperatures in the vicinity of the transition temperature at 1 atmosphere pressure. The closest estimate that can be made of the quantity $\frac{dT}{dP}$ is from the reaction:

$$\text{Borax}(c) = \text{Kernite}(c) + \text{Water}\ (l).$$

For reasons given below in the discussion of the crystal chemistry and kinetics of reaction in the $Na_2B_4O_7$—H_2O system, we write the reaction for the conversion of borax to kernite as:

$$\begin{array}{ccccc} \text{Borax} & = & \text{Kernite} & + & \text{Water} \\ Na_2B_4O_5(OH)_4.8H_2O(c) & = & Na_2B_4O_6(OH)_2.3H_2O(c) & + & 6H_2O(l). \end{array}$$

For this reaction the increase in volume of the system is 31 cc/mole of kernite formed; an increase of approximately 14 percent. For this calculation the following numerical values are assumed: density of borax = 1.715 gm.cm^{-3} density of kernite = 1.908 gm.cm^{-3} (*Dana System of Mineralogy*, 7th ed., 1951, v. II, p. 339 and 336); density of liquid water at 60°C = 0.9832 gm.cm^{-3} (*Handbook of Chemistry and Physics*, 37th ed., 1955, p. 1972). Increase in pressure at a given temperature will therefore favor the formation of borax, i.e., the transition temperature is raised upon increase in pressure.

Menzel and Schulz (1940) have determined ΔH for the reaction

$$\text{Borax}(c) = \text{Kernite}(c) + \text{Water}(g)$$

to be 12.9 Kcal./mole of H_2O formed (at ca. 50°C). We take the heat of vaporization of water to be 10.24 Kcal./mole, at 50°C (*Handbook of Chemistry and Physics*, 37th ed., 1955, p. 2135), and write, then,

$$Na_2B_4O_5(OH)_4.8H_2O(c) = Na_2B_4O_6(OH)_2.3H_2O(c) + 6H_2O(g) \qquad \Delta H = 77.4 \text{ Kcal},$$

and

$$6H_2O(g) = 6H_2O(l) \qquad \Delta H = -61.4 \text{ Kcal}$$

$$Na_2B_4O_5(OH)_4.8H_2O(c) = Na_2B_4O_6(OH)_2.3H_2O(c) + 6H_2O(l) \qquad \Delta H = 16.0 \text{ Kcal}.$$

Using this value of ΔH, the value for ΔV given above, and the transition temperature 331° K in the relationship $\frac{dT}{dP} = \frac{T\Delta V}{\Delta H}$, where P is the total pressure, we calculate the change in transition temperature with pressure:

$$\left.\frac{dT}{dP}\right|_{t=58°C} = 0.015 \text{ deg./bar.}$$

The effect of uniform pressure can be shown graphically by plotting the change of transition temperature with total pressure against various assumed temperature-depth curves (fig. 3). If it is assumed that the geothermal gradient is 2°C per 100 feet and the density of overlying rocks is 2.0 gm.cm^{-3},

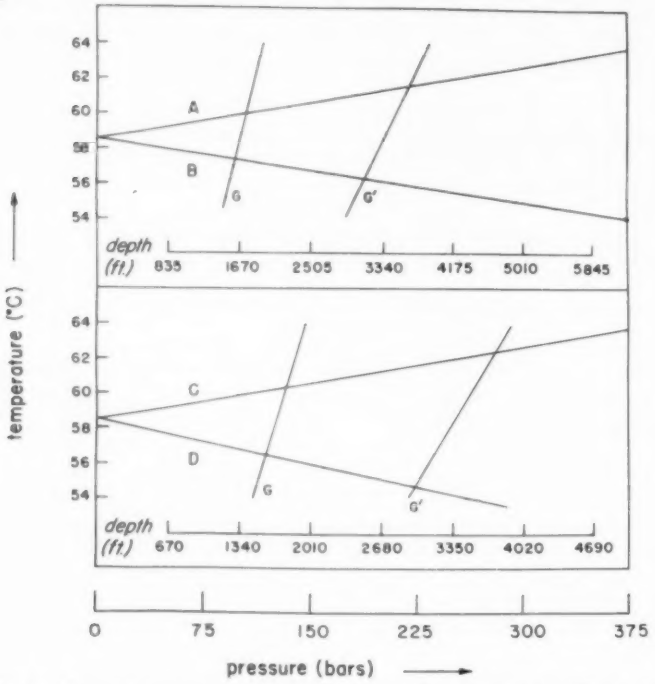

Fig. 3. Pressure-temperature relations for the reaction borax = kernite + water. The lines G and G′ correspond to geothermal gradients of 2°/100 ft., and 1°/100 ft., respectively.

Curve A. Same pressure acting on all phases. Rock density = 2.0 gm.cm^{-3};

$$\left.\frac{dT}{dP}\right|_{58^\circ} = 0.015 \text{ deg./bar.}$$

Curve B. Rock pressure acting on solid phases, hydrostatic pressure acting on liquid phase. $\frac{\text{Rock density}}{\text{Hydrostatic density}} = \frac{2}{1}$;

$$\left.\frac{dT}{dP}\right|_{58^\circ} = -0.012 \text{ deg./bar.}$$

Curce C. Same pressure acting on all phases. Rock density = 2.5 gm.cm^{-3};

$$\left.\frac{dT}{dP}\right|_{58^\circ} = 0.015 \text{ deg./bar.}$$

Curve D. Rock pressure acting on solid phases, hydrostatic pressure acting on liquid phase. $\frac{\text{Rock density}}{\text{Hydrostatic density}} = \frac{2.5}{1}$;

$$\left.\frac{dT}{dP}\right|_{58^\circ} = -0.017 \text{ deg./bar.}$$

then the transition (*at equilibrium*) from borax to kernite would occur at a depth of 1750 feet at a temperature of 60°C. For a gradient of 1°C/100′ and the same density the depth would be 3650 feet and the temperature 62°C; for a density of 2.5 and a gradient of 2°C/100′, the depth would be 1775 feet and the temperature 60°C; for the same density and a gradient of 1°/100′ the depth would be 3750 feet and the temperature 62°C.

THE EFFECT OF NON-UNIFORM PRESSURE

In the preceding section the effect of pressure applied uniformly to both solid and liquid phases of the borate deposit has been considered. As an alternative, the possibility of having the solid phases under lithostatic pressure and the liquid phase under hydrostatic pressure can be treated. In this instance the process would consist of the transition of borax to kernite but with the liquid having essentially free access to the surface.

This situation has been assessed by MacDonald (1953) for the gypsum-anhydrite reaction. Following his procedure:

$$\left.\frac{dP}{dT}\right|_{t=58°C} = \frac{\Delta S}{\Delta V_s + \dfrac{V_{H_2O}}{D}},$$

where ΔV_s = (volume 1 mole kernite) − (volume 1 mole borax), V_{H_2O} is the total volume of water formed, and D is the ratio of the density of the column exerting pressure on the solid phases to that exerting pressure on the liquid phase. For the particular case where the density of the rock column is 2.5 gm.cm^{-3}, and that of the liquid phase is 1.0 gm.cm^{-3}

$$\left.\frac{dT}{dP}\right|_{t=58°C} = -0.017 \text{ deg./bar.}$$

It is seen that the pressure coefficient of transition for non-uniform pressure is of about the same magnitude as that for uniform pressure, but has the opposite sign. The effect of various ratios of rock pressure to hydrostatic pressure on the depth at which transition takes place is shown in figure 3.

THE EFFECT OF DISSOLVED SOLIDS

The effect of dissolved salines on the transition temperature may be readily approximated, when data on the vapor pressure of the saline solutions are available. Consider the extreme case where the transition of borax to kernite takes place in the presence of saturated NaCl solution. The equilibrium vapor pressure of water for the reaction

$$Na_2B_4O_5(OH)_4.8H_2O(c) = Na_2B_4O_6(OH)_2.3H_2O(c) + 6H_2O(g)$$

may be calculated as a function of temperature using the relationship

$$2.303\ n \log \frac{p_2}{p_1} = -\frac{\Delta H}{R}\left(\frac{1}{T_2} - \frac{1}{T_1}\right).$$

For this equation n = 6, and it is assumed that ΔH is constant over the short temperature interval under consideration. We take ΔH = 77.4 Kcal and p = 131 mm at 58.5°C (Menzel and Schulz, 1940). Using the values for the vapor pressure of water over saturated NaCl solution from *International Critical Tables*, 1933, III p. 369, the following table is prepared:

temp.	p_{H_2O} (sat.NaCl)	p_{H_2O} (borax-kernite)
20°C	13.2 mm	10.0 mm
30	23.9	20.8
40	41.6	41.1
50	69.5	78.0

From this table it is seen that the transition temperature is 41°C in saturated NaCl solution, at 1 atmosphere total pressure.

OVERALL EFFECT ON THE REACTION

From the above considerations it is seen that the calculation of the temperature and depth of burial for the equilibrium conversion of borax to kernite depends upon the details of reconstruction of sedimentary environment.

The presence of dissolved salts other than borates would lower the transition temperature, initially, but because of the flushing action of the large volume of water released in the reaction, this initial effect would not be sustained, as has been discussed in detail previously.

The pressure imposed by the environment would probably lie somewhere between purely uniform pressure, and having only lithostatic pressure on the solid phases and hydrostatic pressure on the liquid phase. We consider it unlikely that the pressure in the liquid phase was less than hydrostatic at any time during the precipitation and burial of the deposit. Thus, it seems likely that the bulk of the transition of borax to kernite took place at about 58° ± 5°, and at a depth of 2500 ± 500 feet.

The sporadic nature of the occurrence of kernite masses shows that some overshooting of the equilibrium temperature probably occurred. The overshoot is difficult to estimate, but may not have been large, for Schaller (oral communication) has synthesized kernite from slowly evaporated solutions at 92°C in the laboratory without seeding. The natural heating process was much slower.

CRYSTAL CHEMISTRY AND KINETICS OF REACTION

Consideration of the known crystal structure of borax (Morimoto, 1956), and the probable relationships between this structure and those of tincalconite and kernite, sheds considerable light on the physical chemistry of the $Na_2B_4O_7$—H_2O system. Morimoto has shown that borax crystals contain the polyanion $[B_4O_5(OH)_4]^{-2}$. Thus, the formula for borax is correctly written as $Na_2B_4O_5(OH)_4.8H_2O$, i.e., two of the ten water molecules per formula unit are present in the crystal in the form of hydroxyl groups, and eight are present as actual H_2O molecules. Tincalconite and borax rapidly and reversibly convert to one another upon appropriate changes in humidity, indicating that only small activation energies are involved in the transformation. This fact suggests that tincalconite contains the same $[B_4O_5(OH)_4]^{-2}$ polyions as does borax and that the correct formulation of tincalconite is $Na_2B_4O_5(OH)_4.3H_2O$ (Christ and Clark, 1957). Kernite has two perfect cleavages parallel to (001) and (100) and a less perfect one parallel to (101) so that there is a great tendency for crystals of this mineral to break down into a mass of parallel fine

fibers "resembling a mass of tremolite asbestos" (Schaller, 1930). This morphological observation strongly suggests that the crystal structure of kernite is based on some kind of a chain structure, with the chains running parallel to the fiber axis. The difficulty of converting borax or tincalconite into kernite under conditions where kernite is clearly the thermodynamically stable phase are well-known and have been summarized by Kemp (1956). This difficulty indicates that the conversion requires an appreciable activation energy and thus that kernite does not contain the same $[B_4O_5(OH)_4]^{-2}$ polyanions that are present in borax and tincalconite.

The results of previous crystal structure studies of the series $2CaO.3B_2O_3.xH_2O$ can be used to further clarify the situation. In this series, for colemanite $x = 5$, for meyerhofferite $x = 7$, for a synthetic compound $x = 9$, and for inyoite $x = 13$. Inyoite (Clark, 1959), the synthetic (Clark and Christ, 1957), and meyerhofferite (Christ and Clark, 1956) all contain discrete polyanions of composition $[B_3O_3(OH)_5]^{-2}$. Thus, inyoite is $CaB_3O_3(OH)_5.4H_2O$, the synthetic is $CaB_3O_3(OH)_5.2H_2O$, and meyerhofferite is $CaB_3O_3(OH)_5.H_2O$. On the other hand, colemanite contains infinite chains having the composition $[B_3O_4(OH)_3]_n^{-2n}$ and the formula of colemanite is $CaB_3O_4(OH)_3.H_2O$ (Christ, Clark, and Evans, 1958). The relationship between the isolated polyanions of the higher hydrates and the chains of colemanite can be expressed schematically by the polymerization reaction:

$$n[B_3O_3(OH)_5]^{-2} = [B_3O_4(OH)_3]_n^{-2n} + nH_2O.$$

The $[B_3O_3(OH)_5]^{-2}$ polyanion is composed of two boron-oxygen tetrahedra and a boron-oxygen triangle linked at corners to form a ring. The $[B_4O_5(OH)_4]^{-2}$ polyanion of borax may be derived stereochemically from the $[B_3O_3(OH)_5]^{-2}$ polyanion simply by adding a $BO_2(OH)$ triangle. The polyanions found in meyerhofferite and borax, and the chains found in colemanite, are shown in figure 4.

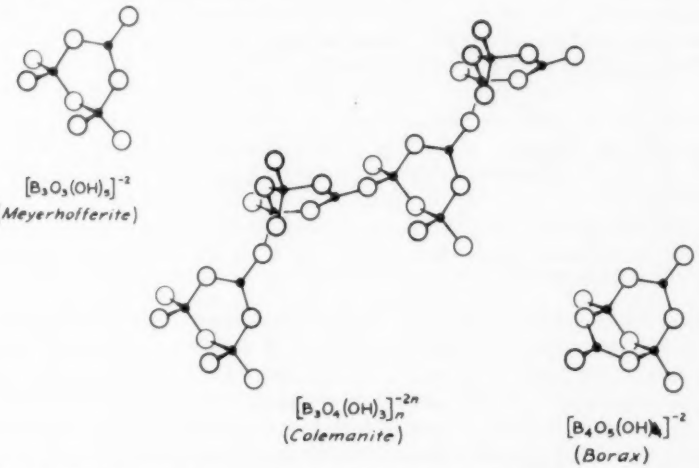

Fig. 4. The structural elements found in colemanite, in meyerhofferite, and in borax.

Considering all of these facts together it was postulated that kernite contains infinite chains of composition $[B_4O_6(OH)_2]_n^{-2n}$ and has the formula $Na_2B_4O_6(OH)_2.3H_2O$ (Christ and Clark, 1957). Kernite would bear the same relationship to borax as colemanite does to inyoite and the relationship between the polyanions of borax and the chains of kerite would be given schematically by the polymerization reaction:

$$n[B_4O_5(OH)_4]^{-2} = [B_4O_6(OH)_2]_n^{-2n} + nH_2O.$$

The formula for borax is correctly written as $Na_2B_4O_5(OH)_4.8H_2O$, and those for tincalconite and kernite are very probably then $Na_2B_4O_5(OH)_4.3H_2O$, and $Na_2B_4O_6(OH)_2.3H_2O$, respectively. If these formulas are accepted the apparently anomalous solubility-temperature relations for the borax-tincalconite-kernite-water system are readily explained. The solubility curves for the three minerals, taken from Menzel and Schulz (1940), were shown in figure 2. As has been pointed out by these authors, ordinarily a system consisting of a higher, an intermediate, and a lower hydrate would yield a solubility-temperature diagram of the kind shown in figure 5. If borax, tincalconite, and kernite are considered to be the 10-, the 5-, and the 4- hydrate, respectively, the solubility curve should be of the type shown in this latter

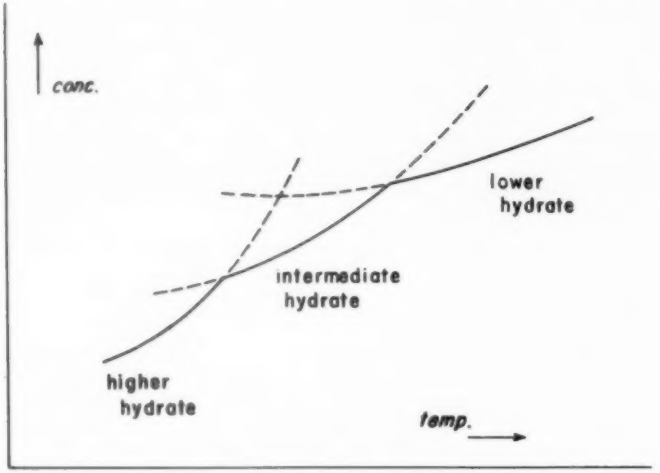

Fig. 5. Schematic drawing of a common type of solubility-temperature diagram for a family of three hydrates. Compare with fig. 2.

figure. On the other hand if, as we have postulated, both tincalconite and kernite contain $3H_2O$ of hydration per formula unit, then no anomaly exists. Moreover, the higher solubility of tincalonite, with respect to kernite, at a given temperature, and the persistence of tincalconite in equilibrium with its saturated solution over the temperature range in which kernite is the truly stable phase (Kemp, 1957) are in accordance with the postulation of discrete polyions in tincalconite and chains in kernite.

Another deduction of considerable interest may be drawn from the above considerations. Since the correct formulas for inyoite, the synthetic, meyerhofferite, and colemanite are known, it is possible to construct a schematic solubility-temperature diagram for the $2CaO.3B_2O_3.xH_2O$—H_2O system, even in the absence of experimental solubility measurements. It seems most likely that such a diagram would be analogous to that of the sodium tetraborate-water system, and would in fact provide a scheme for guiding the experimental measurements, particularly in those regions in which the known crystal chemistry would predict that metastability and nonequilibrium conditions might prevail. Additional support for the reasonableness of such a schematic construction stems from the fact that colemanite and meyerhofferite are known to be secondary minerals, and further both of these minerals occur as pseudomorphs after inyoite.

SUMMARY AND CONCLUSIONS

Relations among the sodium borate hydrates at the Kramer deposit, when considered in conjunction with laboratory work and crystal structure considerations, lead to the following interpretations:

1) The deposit could reasonably have been a primary chemical precipitate of borax at the floor of a saline lake.

2) Deposition of later sediments resulted in an increase in temperature and pressure, and kernite formed in the lower part of the deposit by dehydration of borax. The irregular distribution of kernite is explained by the difficulty of forming the polymerized kernite crystal structure, so that the reaction borax→kernite is extremely slow, and the specific site of formation may depend on "accidents" of nucleation.

3) The abundance of kernite in the lower part of the deposit, and its absence in the upper part suggest that burial was just sufficient to exceed temperature-pressure-composition conditions for the borax-kernite reaction. The temperature of this transition, even allowing for variable salt effects and possible differences in overburden density and thickness, falls in the range 53°-63°C. An overburden thickness of about 2500 feet is suggested.

4) The dehydration of borax to tincalconite on the walls of mine openings, and the rehydration of tincalconite to borax in sealed workings, also fits into what is known of reaction rates and crystal structures. The dry desert air entering the mine has a humidity in equilibrium with kernite or tincalconite (less than 50 percent), but because the rate of formation of kernite is infinitely slow at room temperature, the easily crystallized tincalconite forms. Again the complex polymerization necessary to form kernite makes it difficult to decide whether kernite or tincalconite is the equilibrium species. The persistence of kernite under surface conditions is also explained on the basis of kinetic factors. Where workings have been sealed, and the relative humidity is controlled by water seepage, it would be predicted that the relative humidity would first rise to the equilibrium value for borax-tincalconite and, if sufficient water were available, would be maintained at that value until all tincalconite was converted to borax. Subsequently the humidity would rise to that of a solution saturated with borax.

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REFERENCES

Blasdale, Walter C., and Slansky, Cyril M., 1939, The solubility curves of boric acid and the borates of sodium: Am. Chem. Soc. Jour., v. 61, p. 917-920.

Christ, C. L., and Clark, Joan R., 1957, The nature of the polyions in some borate minerals: Geol. Soc. America Bull., v. 68, p. 1708.

Christ, C. L., and Clark, Joan R., 1956, The structure of meyerhofferite, $2CaO.3B_2O_3.7H_2O$, a $P\bar{1}$ crystal, determined by the direct method of Hauptman and Karle: Acta Cryst., v. 9, p. 830.

Christ, C. L., Clark, Joan, R., and Evans, H. T., Jr., 1958, Studies of borate minerals (III): Crystal structure of colemanite, $CaB_3O_4(OH)_3.H_2O$: Acta Cryst., v. 11, p. 761-770.

Clark, Joan R., 1959, Studies of borate minerals (IV): Crystal structure of inyoite, $CaB_3O_3(OH)_5.4H_2O$: Acta Cryst., v. 12, p. 162-170.

Clark, Joan R., and Christ, C. L., 1957, The crystal structure of $2CaO.3B_2O_3.9H_2O$: Acta Cryst., v. 10, p. 776-777.

Gale, Hoyt, S., 1946, Geology of the Kramer borate district, Kern County, California: Calif. Jour. Mines and Geol., State Mineral. Rept., v. 42, no. 4, p. 325-378.

Kemp, Peter H., 1956, The chemistry of borates, part I: Borax Consolidated Ltd., London, p. 31-40.

MacDonald, Gordon, J. F., 1953, Anhydrite-gypsum equilibrium relations: AM. JOUR. SCI., v. 251, p. 884-898.

Menzel, Heinrich, 1935, Zur kenntnis der Borsäuren und borsauren Alkalisalze. IX. Das system $Na_2B_4O_7$—H_2O: Zeitschr. anorg. allg. Chem., Band 224, p. 1-22.

Menzel, Heinrich, and Schulz, Hans, 1940, Zur kenntnis der Borsäuren und borsauren Alkalisalze. X. Der Kernit (Rasorit) $Na_2B_4O_7.4H_2O$: Zeitschr. anorg. allg. Chem., Band 245, p. 157-220.

Morimoto, Nobuo, 1956, The crystal structure of borax: Mineral. Jour. (Japan), v. 2, no. 1 p. 1-18.

Muessig, Siegfried, and Allen, Robert D., 1957, Ezcurrite ($2Na_2O.5B_2O_3.7H_2O$), a new sodium borate from Argentina: occurrence, mineralogy, and associated minerals: Econ. Geol., v. 52, no. 4, p. 426-437.

Muessig, Siegfried, and Allen, Robert D., 1957, The hydration of kernite ($Na_2B_4O_7.4H_2O$): Am. Mineralogist, v. 42, p. 699-701.

Schaller, Waldemar, T., 1930, Borate minerals from the Kramer district, Mohave Desert, California: U. S. Geol. Survey Prof. Paper 158-I, p. 137-170.

Taylor, Hugh S. (Editor), 1925, A treatise on physical chemistry, vol. II: New York, D. van Nostrand.

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FLUTE- AND GROOVE-CASTS IN THE PREALPINE FLYSCH, SWITZERLAND

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ABSTRACT. Flute casts and groove casts are substratal structures common in Flysch sequences. A study of such structures in the Swiss Flysch indicates: (1) Flute casts are more common in poorly graded or nongraded turbidity current deposits, and groove casts are more common in graded deposits. (2) Flute casts indicate the direction of an earlier stage of turbidity flow than that indicated by groove casts. (3) Multiple sets of groove casts on the undersurface of a sandstone are more common than multiple sets of flute casts. Flute casts and groove casts are different in origin. Flutes are subconical depressions produced where the turbulent water of a turbidity current impinges directly on a muddy sea bottom. Grooves are incised on a muddy sea bottom by the debris of a fluidized sediment mass which is believed to assume laminar motion under a turbidity current. Subsequent filling of the flutes and grooves by sand and silt forms flute casts and groove casts.

Groove casts and flute casts have been observed on the bottom of sandstones from Flysch or Flysch-like sequences of different ages in different parts of the world (see table 1). Turbidity currents are believed to play an important role in the deposition of those sequences which also exhibit many other sedimentary structures such as graded bedding, regular interbedding of coarse and fine clastics, load casts, pull-aparts, convolute bedding, cross-lamination, etc.

Groove casts (fig. 1) are parallel straight ridges on the undersurface of a bed, which is commonly a sandstone. The relief of the ridges, representing the depth of the original groves, ranges from less than 1mm to several centimeters. These straight ridges or lines extend for meters, and they show definite preferred orientation. Groove casts include drag marks and slide marks (Kuenen, 1957). Drag marks are believed to have been grooves cut into a mud layer by the debris carried by a turbidity current. Sand or silt filled in the sculptured grooves, and they are now preserved as casts on the bottom of overlying sandstone layers (Crowell, 1955). Slide marks, on the other hand, are believed to have been scratches left on or in a mud layer after subaqueous sliding or slumping. They, too, are commonly preserved on the bottom of overlying slumped beds (Kuenen and Sanders, 1956; Kuenen, 1957). Slide marks are comparatively rare.

Flute casts (fig. 2) are sharp conical welts on the undersurface of a bed. "One end of the ridge is rounded or bulbous, and the other flares out to merge gradually with the striated bottom surface of the sandstone layer. Flute casts range in size from a few millimeters in width and about a centimeter in length up to more than 10 cm by 40 cm. The place of maximum relief is usually a short distance from the rounded 'beak,' which rises steeply from the flat undersurface with an overhang . . . Flute casts are useful in giving direction of flow along the lineation—the beaked or steep end is upstream" (Crowell, 1955, p. 1360). Flute casts have been known by a number of other names, including "lobate rillmarks" (Clarke, 1918), "Wülste" or "Fliesswülste" (Gürich, 1933; Kraus, 1935), "Strömungs-Marken" (Rücklin, 1938), "flow markings" (Rich, 1950; Kuenen and Carozzi, 1953; Kuenen and Sanders, 1956) "Flow-roll markings" (Rich, 1950), and "Spatulate casts" (Pettijohn, 1957, pl. 2 and

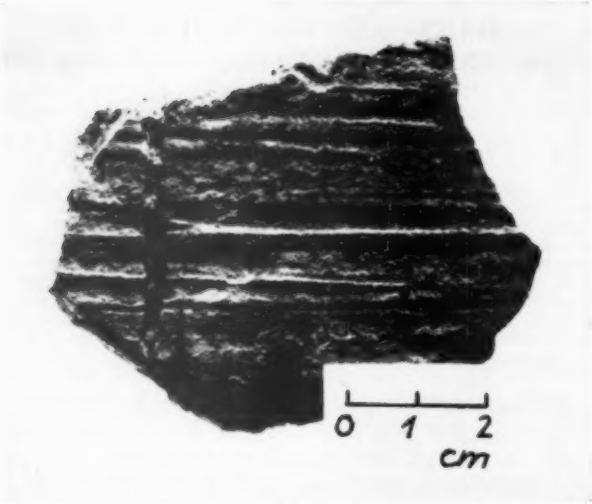

Fig. 1. Groove casts from Schlieren Flysch, southwest of Lake Lucern, Switzerland.

p. 181).[1] As these casts result from filling by sand or silt of lobate incisions on a muddy substratum which are produced by the excavation or fluting of sediment-laden currents, the term "flute casts," first suggested by Crowell (1955), has been recommended for general use (Prentice, 1956; Kuenen and Prentice, 1957).

Groove casts and flute casts are two of the most common directional sedimentary structures on the bottom of a sandstone deposited by turbidity currents. Both structures have been found on the bottom of the same bed (Pettijohn, 1957, pl. 3). However, during the course of study of the Alpine Flysch formations of the Swiss Alps, it became apparent to the writer that flute casts and groove casts do not always occur together. The percentage occurrence of groove casts, giant flute casts, and flute casts in the various Ultrahelvetic Flysch formations has been computed. As shown by table 2, flute casts are twice as common as groove casts in the Wäggital Flysch, in the Basal Schlieren Flysch, and in the Leissigen Flysch, whereas groove casts predominate in the Upper and Middle Schlieren Flysch; both types are equally common in the Lower Schlieren Flysch, which also contains coarse clastic beds with giant flute casts.[2]

[1] Flow casts of Schrock (1948, p. 156, figs. 117 and 118c) resemble what have been described as load casts more than flute casts (Kuenen and Prentice, 1957). Perhaps flute casts were grouped under the "scour and fill" structures of Schrock (1948, p. 230). "Gefliess-Marken" (Richter, 1935; Hantzschel, 1935; Kiesslinger, 1937) resemble what have been described as "flow ridges" on the upper surface of a sandstone bed (Glaessner, 1958), which are genetically and morphologically quite distinct from flute casts (Rücklin, 1938).

[2] Wäggital Flysch and Schlieren Flysch are Flysch formations in the Ultrahelvetic zone of the Swiss Alps. The former is Late Cretaceous (Leupold, 1942), and the latter ranges from Maestrichtian to Lower Eocene (Schaub, 1951). The paleogeographic significance of directional sedimentary structures in these Flysch formations will be discussed in a forthcoming paper (Hsu, in press).

TABLE 1
Occurrences of Groove Casts and Flute Casts

Age	Formation	Locality	References
Precambrian	Marinoan	Woomera, South Australia	Glaessner, 1958
Precambrian	Flagstone	Pretoria, South Africa	Gürich, 1933
Cambrian		Harlech Dome, Wales	Kopstein, 1954
Ordovician		Victoria, Australia	Hills and Thomas, 1954
Ordovician	Martinsburg	Pennsylvania	Pettijohn, personal communication
Silurian		Gaspe, Canada	Clarke, 1918
Silurian		Southern Uplands, Scotland	Kuenen, 1957
Silurian	Aberstwyth Grit	Wales	Rich, 1950; Prentice, 1956
Devonian	Portage, Chemung	New York	Hall, 1843; Clarke, 1918; Kuenen, 1956; Pettijohn, 1958
Mississippian	Bedford	Ohio, Kentucky	Rich, 1950; Rich and Wilson, 1950
Mississippian	Horton	Cape Breton Island, Nova Scotia	Shrock, 1948
Mississippian	Kulm Graywacke	Harz Mts., Germany	Kuenen and Sanders, 1956
Pennsylvanian	Jackfork	Ouachita Mts., Oklahoma	Rich, 1950; Bokman, 1953
Pennsylvanian	Flözleeres Graywacke	Harz Mts., Germany	Kuenen and Sanders, 1956
Triassic	Muschelkalk	Saarland, Germany	Rücklin, 1938
Jurassic	Aalenian Flysch	Carparthians, Poland	Birkenmayer, 1957
Cretaceous		Ventura Basin, California	Jennings & Troxel, 1954
Cretaceous	Wäggital Flysch	Schwyz, Swiss Alps	Kraus, 1932; Hsu, 1958, 1959
Cretaceous	Helminthoid Flysch	Maritime Alps, France & Italy	Kuenen *et al.*, 1957
Cretaceous	Sievering Beds	Vienna, Austrian Alps	Glaessner, 1958
Early Tertiary	Ultrahelvetic Flysch	Voirons	Kuenen & Carozzi, 1953
Early Tertiary	Gurnigel Flysch	External Prealps, Switzerland	Crowell, 1955; Hsu, 1958, 1959
Early Tertiary	Schlieren Flysch	Prealps, Central Switzerland	Hsu, 1958, 1959
Early Tertiary	Ultrahelvetic Flysch	Internal Prealps, Switzerland	Hsu, 1959
Early Tertiary	Greifenstein	Austrian Alps	Glaessner, 1958
Early Tertiary	Autochtonous Flysch	Glarus Alps, Switzerland	Hsu, 1959
Early Tertiary	Autochtonous Flysch	Maritime Alps, France & Italy	Kuenen *et al.*, 1957
Early Tertiary	Flysch Sandstones	Northern Appennines, Italy	Kuenen, 1957; ten Haaf, work in prog.
Pliocene	Pico*	Ventura Basin, California	Crowell, 1958

*As a rule, groove casts are found on the bottom of sandstone beds in well-indurated sedimentary sequences. However, grooves are preserved on the top of indurated shales of the Pico formation; these grooves are exposed after the removal by weathering and erosion of the overlying poorly consolidated sands.

Fig. 2. Flute casts from Leissigen Flysch, south of Thurnersee, Switzerland.

TABLE 2

Percentage Occurrences of Groove Casts, Giant Flute Casts, and Flute Casts in Various Ultrahelvetic Flysch Formations of the Swiss Alps

	No. of Measurements	Groove Casts	Giant Flute Casts	Flute Casts
Wäggital Flysch	37	32%	0%	68%
Upper and Middle Schlieren Flysch	124	85%	1%	14%
Lower Schlieren Flysch	42	58%	10%	32%
Basal Schlieren Flysch	36	30%	0%	70%
Leissigen Flysch	28	21%	0%	79%
Total	267			

Although all these Flysch formations are characterized by a regular interbedding of shales and coarser clastics, subtle differences in lithology were detected by field examinations. The coarser clastic interbeds of the Middle and Upper Schlieren Flysch are almost invariably sandstones with distinct graded bedding, grading from medium or coarse grained sand at the base to silts on the top. These sandstones are rather poor in interstitial clay, and some are actually friable upon weathering. In contrast, the coarser clastic interbeds of the Basal Schlieren Flysch, the Wäggital Flysch, and the Leissigen Flysch are predominantly poorly sorted sandstones with considerable interstitial clay and calcareous cement; some, in fact, should be properly classified as calcareous siltstones. Graded bedding is present, but it is not as common here as in the Middle and Upper Schlieren Flysch. Flute casts are quite commonly found on the bottom of calcareous siltstones which show no apparent graded bedding.

Different still from both the Basal Schlieren and the Middle and Upper Schlieren is the Lower Schlieren Flysch, which contains conglomerates and coarse grained sandstones. Giant flute casts have been found on the bottom of these coarse interbeds which show poor, if any, graded bedding.

The degree of graded bedding has been suggested as a criterion in determining the distance of transport (Kuenen and Menard, 1952). At an area near the source, the highly turbulent turbidity currents arrive at the site of deposition without having had time for velocity differentiation; turbid waters passing through the site at successive instants have comparable velocities and, therefore, comparable transporting powers. Clasts settled from such a current of constant velocity should be of about the same size, and therefore, graded bedding should not be developed near the source. On the other hand, in an area far away from the source of flows, the faster-traveling current, capable of carrying the coarsest particles, forms the front of the flow which arives first and deposits the coarsest sediment at the bottom of a graded bed; the slower-traveling current, not capable of carrying coarse particles, forms the tail of the flow which arrives last and deposits the finest sediments at the top of a graded bed.

Flute casts are more common on the bottom of a muddy and calcareous sandstones and siltstones of the Wäggital, the Basal Schlieren, and the Leissigen Flysch which commonly show indistinct or no graded bedding; giant flute casts are common only on the bottom of coarse conglomerates of the Lower Schlieren Flysch. Thus, flute casts are probably formed at an earlier stage of transport when the passing turbidity currents are still highly turbulent. In contrast, groove casts are more common on the bottom of the Middle and Upper Schlieren sandstones which show distinct graded bedding and which have been somewhat sorted. These graded beds were probably deposited at a later stage when the turbulence of the depositing currents had subsided somewhat during the course of transport. Such inferences are supported by the fact that the current direction indicated by the flute casts and that by the groove casts in the Swiss Flysch are on the average different. The direction indicated by the flute casts is in general more nearly parallel to the direction of ancient submarine canyons down the slope of a Flysch basin; the direction indicated by the groove casts in in general more nearly parallel to the direction of the bathymetric axis of the basin (Hsu, 1959). It has generally been agreed that flutings of the substratum are produced by the turbulence of turbidity currents (Rich, 1950; Prentice, 1956; Kuenen, 1957). Rich (1950) compared this process to the fluting of hard rock surfaces by swift, sand-laden river waters (Maxson and Campbell, 1935; Maxson, 1940). Grooves or drag marks, on the other hand, are not produced by the turbid water itself, but by the sharp-edged debris (shell fragments or other angular or subangular grains) which is dragged along the bottom by the turbidity currents. In some instances, straight grooves extend for several meters (Hsu, 1959; Crowell, 1955, fig. 2). Thus, the particles responsible for grooving must have traveled in a straight line for a considerable distance along the bottom. Such a mode of particulate motion suggests laminar motion of the enclosing fluid. The presence of groove casts suggests, then, that a laminar flow might exist, at least temporarily, at the bottom of some turbidity currents.

Based on theoretical speculations, the following hypothesis is offered to explain the different distribution and the contrasting origins of these two forms of substratal structures: Flutes are produced by a turbidity current which carries most of its load in suspension; the turbulent, sediment-laden water of the current impinged upon a muddy substratum results in excavation of subconical depressions—a process known as fluting. Subsequent filling of those depressions by sand or silt forms flute casts. Drag grooves are produced by a moving sediment-water mass which has already decreased its velocity considerably so that its coarsest load is being deposited. The main body of the mass still moves as a turbidity current which carries its load in traction and suspension. Meanwhile, the coarsest sedimentary particles begin to settle out of suspension and to congregate at the bottom of the turbidity flow. A thin sheet of fluidized sediment mass of high viscosity is thus formed between the turbidity flow and the muddy sea bottom. This mass is forced to continue its forward motion, for a short distance at least, because of the strong shearing stress being exerted on the mass by the turbidity flow above. As the Reynolds number is small for a fluid of high viscosity, the fluidized sediment mass at the bottom of a turbidity current moving at low velocity must therefore assume laminar motion. Sedimentary particles dragged along by the laminar flow cut straight grooves in a muddy substratum. Subsequent filling of those grooves by sand or silt forms drag marks (a form of groove casts).

The working hypothesis offered herein might also explain the fact that two or three sets of groove casts are common occurrences (Crowell, 1958; Kuenen and ten Haaf, 1958), but that two or three sets of flute casts are rarely found (Hsu, 1959). Sheet sands deposited by laminar motion at the base of a turbidity current could easily be picked up by subsequent currents flowing at some different direction, resulting in the formation of multiple sets of grooves before final burial. In contrast, the highly turbulent currents that produce flute casts also deposit poorly sorted coarse or muddy sediments which are more difficult to erode than sheet sands (Hjulström, 1939). The flutings, once formed, are thus buried before another set of flutings can be incised on the same muddy substratum.

Another corollary of the working hypothesis is that the graded sands might not settle directly from the turbulent water of turbidity currents; in some instances, when particles of a certain size become too coarse to remain in suspension, they settle near the bottom of the moving sediment-water mass and are then dragged along for some distance by laminar flow before being permanently deposited on the bottom. Such a mechanism of deposition of graded beds should explain (1) why laminations have been observed in some Flysch sandstones with graded bedding, and (2) why some sandstones deposited by turbidity currents are relatively free of interstitial clay.

In conclusion, the writer wishes to emphasize that very little is known about the mechanism of turbidity current deposition. He offers the idea that laminar motion might be developed at the base of some turbidity currents in order to explain (1) the contrasting origins of flute casts and groove casts, (2) the difference in occurrence of flute casts and groove casts, (3) the difference in orientation of flute casts and groove casts, (4) the common absence of multi-

ple sets of flute casts on the bottom of the same bed and the common occurrence of multpile sets of groove casts on the bottom of the same bed, (5) the laminatins of some sands deposited by turbidity currents, and (6) the lack of interstitial clay in some sands deposited by turbidity currents. It is hoped, that future workers on turbidity current deposits will make an effort to collect more pertinent data to test the working hypothesis offered in this paper.

References

Birkenmayer, K., 1957, Sedimentary characteristics of the Flysch-Aalenian in the Piening Klippen belt (Central Carparthians): Acad. Polonaise Sci. Bull., Cl. 3, v. 5, p. 451-456.

Bokman, J., 1953, Lithology and Petrology of the Stanley and Jackfork formations: Jour. Geology, v. 61, p. 152-170.

Clarke, J. M., 1918, Strand and undertow markings of Upper Devonian time as indications of the prevailing climate: New York State Mus. Bull. 196, p. 199-238.

Crowell, J. C., 1955, Directional-current structures from the Prealpine Flysch, Switzerland: Geol. Soc. America Bull., v. 66, p. 1351-1384.

————, 1958, Sole markings of graded graywacke beds: A discussion: Jour. Geology, v. 66, p. 333-335.

————, 1958, Directional-current structures in restricted geosynclinal basins: Lecture given at American Petroleum Institute Primary Structures in Sediments Symposium, September 25-27, 1958, Houston, Texas.

Glaessner, M. F., 1958, Sedimentary flow structures on bedding planes: Jour. Geology, v. 66, p. 1-7.

Gürich, G., 1933, Schrägschichtungsbogen und zapfenartige Fliesswülste im "Flagstone von Pretoria": Z. deutsch. geol. Ges., v. 85, p. 652-663.

Hall, James, 1843, Geology of New York, Part IV (Survey of the Fourth Geological District): Albany, N. Y., 525 p.

Häntzschel, W., 1935, Fossile Schrägschichtungsbogen, "Fliesswülste" und Rieselmarken aus dem Nama- Transvaal- System (Südafrika) und ihre rezenten Gegenstücke: Senckenbergiana, v. 17, p. 167-177.

Hills, E. S., and Thomas, D. E., 1954, Turbidity currents and graptolite facies in Victoria: Jour. Geol. Soc. Australia, v. 1, p. 119-133.

Hjulström, Filip, 1939, Transportation of detritus by moving water *in* Recent marine sediments (P. D. Trask, ed.): London, Thomas Murby and Co., p. 5-31.

Hsu, K. J., 1958, Paleocurrent structures in Ultrahelvetic Flysch of Swiss Alps and their paleogeographic and paleotectonic significances, (abstract): Geol. Soc. America Bull., v. 69, p. 1588-1589.

————, in press, Directional Sedimentary Structures in Ultrahelvetic Flysch, Switzerland.

Jennings, C. W., and Troxel, B. W., 1954, Ventura Basin *in* Geology of Southern California (R. H. Jahns, ed.): Calif. Div. Mines Bull. 170, Geologic Guide No. 2, 63 p.

Kiesslinger, Alois, 1937, Eine boden-physikalische Betrachtung derGefliess-Marken (Fliesswülste): Senckenbergiana, v. 19, p. 127-138.

Kopstein, F. P. H. W., 1954, Graded bedding of the Harlech Dome: Geol. Inst., Groningen, Netherlands, Publ. 81, 97 p.

Kraus, Ernst, 1932, Ueber den Schweizer Flysch: Eclogae geol. Helv., v. 25, p. 39-129.

————, 1935, Ueber Sandsteinwülste: Z. deutsch. geol. Ges., v. 87, p. 354-360.

Kuenen, Ph. H., 1956, Problematic origin of the Naples rocks around Ithaca, N. Y.: Geol. en Mijnbouw, v. 18, p. 277-283.

————, 1957, Sole markings of graded graywacke beds: Jour. Geology, v. 65, p. 231-258.

Kuenen, Ph. H., and Carozzi, A., 1953, Turbidity currents and sliding in geosynclinal basins of the Alps: Jour. Geology, v. 61, p. 363-373.

Kuenen, Ph. H., Faure-Muret, A., Lanteaume, M., and Fallot, P., 1957, Observations sur les Flyschs des Alpes maritimes françaises et italiennes. Sédimentologie, variation de faciès: Soc. géol. France Bull., 6th ser., v. 7, p. 11-26.

Kuenen, Ph. H., and Menard, H. W., 1952, Turbidity currents, graded and nongraded deposits: Jour. Sed. Petrology, v. 22, p. 83-96.

Kuenen, Ph. H., and Prentice, J. E., 1957, Flow-markings and load-casts: Geol. Mag., v. 94, p. 173-174.

Kuenen, Ph. H., and Sanders, J., 1956, Sedimentation phenomena in Kulm and Flözleeres graywackes, Sauerland and Oberharz, Germany: Am. Jour. Sci., v. 254, p. 649-671.

Kuenen, Ph. H., and ten Haaf, E., 1958, Sole markings of graded graywacke beds: a reply: Jour. Geology, v. 66, p. 335-337.

Leupold, W., 1942, Neue Beobachtungen zur Gliederung der Flyschbildungen der Alpen zwischen Reuss und Rhein: Eclogae geol. Helv., v. 35, p. 247-291.

Maxson, J. H., 1940, Fluting and faceting of rock fragments: Jour. Geology, v. 48, p. 717-751.

Maxson, J. H., and Campbell, Ian, 1935, Stream fluting and stream erosion: Jour. Geology, v. 43, p. 729-744.

Pettijohn, F. J., 1957, Sedimentary rocks, 2nd ed.: New York, Harper and Brothers, 718 p.

————, 1958, Primary structures and paleocurrents: Lecture given at American Petroleum Institute Primary Structure in Sediments Symposium, September 25-27, 1958, Houston, Texas.

Prentice, J. E., 1956, The interpretation of flow-markings and load-casts: Geol. Mag., v. 93, p. 393-400.

Rich, J. L., 1950, Flow markings, groovings, and intra-stratal crumplings as criteria for recognition of slope deposits, with illustrations from Silurian rocks of Wales: Am. Assoc. Petroleum Geologists Bull., v. 34, p. 717-741.

Rich, J. L., and Wilson, W. J., 1950, Paleographic and stratigraphic significance of subaqueous flow markings in the Lower Mississippian strata of south-central Ohio and adjacent parts of Kentucky (abstract): Geol. Soc. America Bull., v. 61, p. 1496.

Richter, R., 1935, Marken und Spuren im Hunsrück-Schiefer, I, Gefliess-Marken: Senckengergiana, v. 17, p. 244-263.

Rücklin, Hans, 1938, Strömungs-Marken im unteren Muschlelkalk des Saarlandes: Senckenbergiana, v. 20, p. 94-114.

Schaub, Hans, 1951, Stratigraphie und Mikropaläontologie des Schlierenflysches, mit besonderer Berüsichtigung der paleocänen und untereocänen Nummuliten und Assilinen: Schweiz. Pal. Abh., v. 68, p. 1-222.

Shrock, R. R., 1948, Sequence in layered rocks: New York, McGraw-Hill Book Co., 507 p.

[American Journal of Science, Vol. 257, Summer 1959, P. 537-538]

DATING CHAOS JUMBLES, AN AVALANCHE-DEPOSIT IN LASSEN VOLCANIC NATIONAL PARK

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ABSTRACT. Counts of tree-rings were employed to date a major avalanche deposit. This technique required the evaluation of assumptions as to delays in seeding and errors in coring. The data suggest a date of origin between 1684 and 1696.

The Chaos Jumbles in the northwestern portion of Lassen Volcanic National Park is a huge avalanche deposit with a typical hummocky topography. The deposit is composed of angular dacite fragments and forms an extremely rough terrain. It supports a sparse forest of various conifers with the trees widely spaced and frequently stunted, yellowish, with broken tops and other features indicative of difficult survival. There are, however, few dead or down trees.

In the original major study of this area, Williams (1928) mentioned but one avalanche and estimated its age then as 200 years. There are indications that there have been older avalanches but the techniques employed here were restricted to use only in areas unquestionably of most recent origin.

Cores were taken with a Swedish increment borer, and the trees were plugged with dowling and painted with sealant to prevent insect or fungus access. Each tree was mapped.[1] The coring procedures, although relatively simple under most conditions, proved to be quite troublesome. The trees are bent and crushed by snows during their early years and strong prevailing westerly winds distort their growth still more. Few cores hit the exact heart of a tree so that it was necessary to draw completion circles of the rings of early years and estimate those missing. Five or fewer missing rings can be estimated with moderate accuracy but greater departure from the heart seriously diminishes the value of the core. Attempting another core is often impractical for many of the trees have wood so dense that the strain on the coring tool is excessive. What is more, the irregularities mentioned often make a second try little better than the first.

Cores were oriented in grooves in hardwood (Heath, 1958) and planed to very smooth surfaces so that accurate counts could be made. Since some trees exhibited nearly 100 years' growth to an inch, only excellent surfaces were suitable.

In considering mechanical problems it should be noted finally that the coring tool must be used approximately 15 inches above the tree base thereby eliminating the years taken for this much development. Applying a very conservative estimate, seven years were added to all tree-core totals to offset this factor.

How long after the formation of such an avalanche deposit a suitable seedbed would be provided is a problem that must be judged on the basis of evidence provided elsewhere in the park. Lassen Peak was the source of a major mudflow in 1915. Today, this area is covered with many conifers. In 1957 a second mudflow buried part of this same area. On a permanent quadrat of

[1] Maps are on file at the park.

5000 square feet, three conifer seedlings appeared the following spring and, whether or not they survive, it is clear that invasion can be exceedingly rapid.

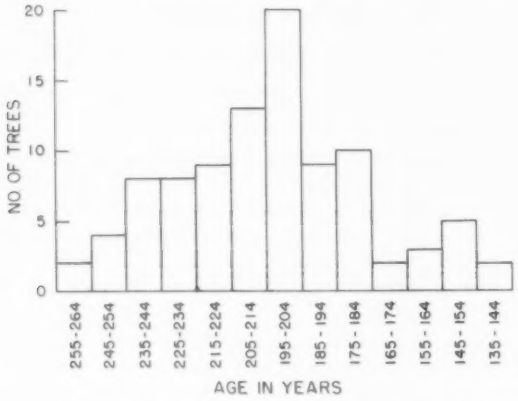

Fig. 1. Distribution of trees by age group.

The 92 satisfactory cores present the distribution shown in figure 1. The increasing number of trees found with decreasing years suggests gradual establishment of the forest. The decrease and erratic pattern of still younger groups is merely evidence of poorer discrimination in selection of trees worth coring. Strictly random coring would increase these younger groups considerably.

On the basis of the foregoing data and assumptions, the recent avalanche of the Chaos Jumbles seems to have originated not less than 263 years ago and probably not more than 275 years ago. Williams' estimate was very close.

Work in the future will be devoted to analysis of the older avalanche deposits.

ACKNOWLEDGMENTS

This work was supported in part by a grant from the American Academy of Arts and Sciences. I should like also to thank the National Park Service, Park Superintendent Freeland, Park Naturalists V. R. Bender and R. Prasil, and District Ranger L. Bodine. Dr. Gordon Macdonald offered many helpful suggestions.

REFERENCES

Heath, J., 1958, Increment core techniques and preventing core shrinkage: Jour. of Forestry, v. 56, p. 909-910.

Williams, H., 1928, A recent volcanic eruption near Lassen Peak, California: Univ. Calif. Publ., Dept. Geol. Sci., v. 17, p. 241-263.

ESSAY REVIEW

Geologisches Kräftespiel und Landformung. Grundsätzlische Erkenntnisse zur Frage junger Gebirgsbildung und Landformung; by ARTHUR WINKLER-HERMADEN. P. xx, 822; 120 figs., 5 plates. Vienna, 1957 (Springer-Verlag, $22.15).—This well written book deals primarily with the problems of the mutual relationship between the geotectonic elements that built mountain chains and landforms in Cenozoic time. The final aim of the author's studies was to obtain a proper perception of the dynamic equilibrium of the endogenic and exogenic processes which form the Earth's surface. Although some conclusions and interpretations of general importance are derived from a relatively restricted area—and from this point of view the inclusive title of the book might be somewhat misleading—the author always keeps his notions within the frame of the particular conditions of that area and their local properties. However, since the part of the Alps discussed is a region of such multiplicity, being to some extent a key area in geologic research, the results presented have general value as well. The author has done research in that region for about forty-five years and is more familiar than most people with all the problems in stratigraphy, geomorphology, geotectonics, and allied topics as well as with the tremendous literature on the subject. For this reason—and also because of the voluminous presentation—it is not easy to compose a proper view.

The enormous overall material is so well organized that the only wish the reader might have is that the bibliography as well as the locality-index were in a form easier to look over. Logical sequence and a consistent derivation of conclusions enforce subdivision of the book into a few main parts.

The first part is by far the shortest one (p. 1-15). It is more or less an introduction in which the author merely points out some general problems and fundamental statements, and the direction of his own inquiries and opinions. Without detailed argumentation, the Cenozoic events in that area (including the greatest part of the Eastern Alps and their extension to the Carpathian Mountains and the Dinaride chain as well as the forelands and basins north, east, and south of the Alps) are first of all subdivided into five phases (main cycles of second category). These are basically identical with Stille's subdivision of the Cretaceous-Alpine Orogenesis (a main cycle of first category) into seven phases. The author, however, accepts only the first four phases of Stille —the Lamarische, Pyrenäische, Savische, and Steirische phases, whereas the latter three phases of Stille are compressed into a last Piocene cycle (phase) in which the present geomorphologic features have been formed (apart from modifications during Quaternary time). Every phase consists of several long-term episodes of intensive orogenic faulting separated by shorter episodes in which epirogenic and the development of the drainage net predominated.

In the second part the author gives a minute delineation of the Late Miocene and Pliocene events; the Pleistocene is treated much more briefly and more or less in outline, since the author has recently presented (Denkschriften der Österreichischen Akademie der Wissenschaften, math.-naturwiss. Klasse, v. 110, Vienna, 1955) a detailed paper referring to this particular subject. The geological processes forming the landscape are delineated as they operated in the various succeeding stages—beginning with the Sarmatian stage of the Miocene—within certain regions or provinces that are considered as

tectonic-geomorphological space units. These are: the Styrian basin and its periphery, the range of the late Tertiary Save faults, the Tertiary province at the Venetian-Friaulian rim and at the flange east of the Southern Alps, the West Pannonian basin, the Interior-Alpine Vienna basin and its mountain border, the northern foreland of the Alps in Salzburg as well as Upper- and Lower-Austria (with particular reference to the geomorphology of the northern Limestone Alps), and finally the Tertiary belt within the Eatsern Alps (with reference to the geomorphologic development of the eastern Central Alps).

Chapters A to E (p. 15-328) deal chiefly with the stratigraphy. In Chapter D the Late Tertiary sequence is discussed and compared with that of Southern, Western, and Eastern Europe, and in Chapter E the problem of the Pliocene-Pleistocene boundary is briefly summarized. Finally the author offers some suggestions for a reasonable new subdivision of the Pliocene. This seems to be necessary especially with regard to the separation of some substages which have been considered in former time as latest Pliocene but now are regarded as belonging to the oldest Pleistocene. The author's argumentation is precisely stated and his opinions are summarized into lucid tables.

In Chapters F to L (p. 329-604) the geologic-geomorphologic development within the various subregions (provinces) is delineated. The author starts his considerations with the Recent (Holocene) and Late Pleistocene terraces and goes back to successively older forms and geomorphological processes.

All these results of regional research are the basis for the third part (Chapters A to J; p. 605-746) in which the author points out mainly consequences and decisions concerning the problems of Cenozoic sequence, tectonism, volcanism, and geomorphology, in addition to some conclusions of more general nature and importance, such as relationships between tectonic and volcanic activity (Chapter C), depth movements within the orogens and their geophysical interpretation (Chapter E), and problems in interpretation of transgressions and regressions (Chapter F).

This third part particularly (the fourth and last part is just a very brief summary) will certainly be a starting point for discussion and further research. There are some very interesting ideas here, which will shed light particularly on many problems in stratigraphic co-ordination and correlation. Thus, for instance, the author shows convincingly that an orogenic episode, represented by a certain fault trace at the southeastern border of the Alps, has produced contemporaneously in the interior only an arch or even simply a gentle uplift. Doubtful, however, seems the statement that mountainous mass-subsidence occurs not only as marginal tilting, or only within certain fields or zones, but also in the form of tremendous inclinations or "downbendings". The author indeed confesses that there is no real proof available for this opinion, but it seems to the reviewer that there is no reason at all for such an assumption, because all cases mentioned show zones of more or less local subsidence within a massif undergoing progressive uplift. The deposition of fine-grained sands along some mountain borders suggests tectonic inactivity rather than such extensive subsidence or "downbending" as supposed by the author.

He emphasizes, certainly rightly, that the step-topography of older mountain systems is primarily the product of a dynamic equilibrium between endo-

genic and exogenic processes. The problem therefore has to be considered mainly as a matter of morphotectonics, but he admits that climatic conditions played an important role also, promoting extensive planations. These were especially favored by deep rock-decomposition and weathering, soil movements, and eluviation beneath the Tertiary vegetative cover. Although such considerations and interpretations appear very subordinated to the tectomorphic viewpoints (brought sometimes too sharply into focus), valuable information and hints are offered concerning quality and quantity of removal by erosion and denudation as well as properties and significance of Tertiary climates as related to landforms. The author gives fairly good additional evidence that climatic conditions from Late Miocene until Late Pliocene were dominantly moist-tropic, but included also several arid episodes or substages. Only towards the end of Pliocene did a transition to moderate climates take place.

Surprising new opinions, however, are stated in the interpretation of Pleistocene geomorphology. According to the author's assertions, only the two youngest terraces, equated with the Würm and Riss glaciation, are deposits developed during cold climates. They are composed by material which has been brought dominantly from the glaciated region but without essential cooperation of periglacial effects. For all the higher terraces an interglacial age is assigned, and their origin is referred to effects of mutual tectonic and eustatic movements.

Apart from the fact that actually the Würm and Riss glaciation is represented not only by the two lowest terraces in that area, the author's conception is, as a whole, not acceptable. It is quite true that some gravel bodies, for instance those of the Drau river where it enters the Prettau field, consist almost entirely of material brought from the glaciated region, but this is absolutely not typical for the complete area and therefore is an unacceptable generalization. As a matter of fact, the author was working in Pleistocene geology chiefly in southeastern Styria, close to the margin of a region which was very probably even during glaciations out of reach of periglacial climatic conditions. It seems to the reviewer therefore that the conclusions were derived from this restricted and typical region. It must be emphasized that there can be no doubt that in almost all other gravel bodies the largest amount of material was brought in from periglacial mass-wasting. This is especially true of the big river terraces along the Danube and its tributaries, because many of them—not only the northern but also some southern ones—had no immediate connection with the glaciated region at all.

Concerning the opinion that the higher terraces are interglacial, it also needs to be pointed out that there is no reason for such an interpretation. On the contrary, all evidence suggests a glacial origin. There is no space here either to delineate all arguments that speak for this latter assertion or to criticize some other statements of the author on this particular subject, because that would be a matter of only local interest. However, it should be mentioned, at least, that without exception all these older gravel terraces show the same features as the younger ones, and all—especially those along the Danube and its southern tributaries—have big boulders near the bases, some even with scratches which suggest that they have been brought in by a certain kind of

ice drifting, as described particularly in some papers by H. Küpper in 1954 and 1955. Some other papers, too, dealing with Pleistocene geology, terraces, etc., seem not to have come to the author's attention.

As a whole the book is very stimulating. It is a very good summary of events during the period from the Late Miocene to the beginning of the Pleistocene of the eastern Alps and adjacent regions, and fills a long-lasting gap in the geological literature. We must be very grateful to the author for bringing order and reason into the multiplicity of problems and facts; it certainly represents hard work over many years.

F. BRANDTNER

REVIEW

Geology of the Great Lakes; by JACK L. HOUGH. P. xviii, 313; 75 figs. Urbana, 1958 (University of Illinois Press, $8.50).—This monograph is the first definitive work on the Great Lakes since the momentous work of Frank Leverett and Frank Taylor forty-five years ago. Within the pages of this book the reader will find a wide variety of subject material ranging from regional geology of the Great Lakes drainage basins to the radiocarbon chronology of the various pre-modern lake stages, presented in a bold and straightforward fashion by an author well qualified for the task.

The book is divided into two parts. Part I contains six chapters: 1. The Modern Great Lakes, 2. The Lake Water, 3. Sediments of the Great Lakes, 4. Preglacial History of the Region, 5. Glacial History of the Region, and 6. Dating the Events of Lake History. Part II is concerned with the details of the history of the lake stages and ends with 26 maps showing the author's conception of the complex sequence of events in the waning ice front, changing lake levels, and shifting lake outlets. A correlation chart of stages in the five lake basins of Superior, Michigan, Huron, Erie, and Ontario is also given. The time scale is based on radiocarbon dates.

Part I of the book is mostly a modern treatment of existing knowledge although the author has some interesting viewpoints which he develops. One of these is the statement on page 34 that " . . . there is no depth that can be assigned to wave base." His argument is based on the fact that coarse lag gravels occur near Chicago at a depth of 50 feet, and that the hull of the tug *Walker* which sank in 65 feet of water near Nicholson's Island in Lake Ontario was " . . . battered to pieces, and . . . 75 tons of coal which had been aboard were washed away." (p. 35). Also, the author points out that the bottom of Lake Michigan in the mid-lake shallow area where depths range from 132 to 200 feet " . . . is somehow kept free of finer-grained sediments at the present time." (p. 35). Many will question Hough's flat conclusion that the limit of wave action is not roughly equal to one-half the wave length. As a matter of fact, Hough himself cites instances on page 32 of undisputed wave lengths in excess of 300 feet on the Great Lakes, which would permit some wave scouring at depths of 150 feet during storms if the wave base = ½ wave length formula were applied.

Another section in Part I which will undoubtedly interest Pleistocene

geologists is the adoption of a Wisconsin time scale (p. 94) in which the Iowan substage is called Shelbyville-Iowan (early Tazewell), and the Tazewell is called the Bloomington (late Tazewell). Also, the Mankato is replaced by the Port Huron (Mankato) and the Cochrane is recognized as having full substage rank.

Part II, History of Lake Stages, is essentially an elaboration on the author's previous work which came out in mimeographed form in 1953 and was entitled, "Final Report on the Project, Pleistocene Chronology of the Great Lakes Region" (Office of Naval Research contract no. N6 ori-07133). Some new and radical changes in the old Leverett and Taylor chronology are proposed, some of which are well documented and completely accepted by geologists actively engaged in research and thoroughly familiar with the details of Great Lakes Pleistocene geology. Among such revisions the reader will find a thorough account of the Chippewa and Stanley low-water stages which the author named and documented on the basis of his own research on the lake bottom cores in the Lake Michigan Basin.

In the category of less thoroughly documented revisions the reader will be interested in, though not necessarily convinced by Hough's correlation of the Grassmere, Lundy, and Algonquin stages in the Huron-Erie basins with the Glenwood, Calumet, Toleston stages in the Lake Michigan Basin. Since the Grassmere and Glenwood beaches are both at 620′ A.T., and the Algonquin and Toleston beaches are at 605′ A.T., the author concludes on page 159 that: "It seems beyond the limits of probability that three lake stages in the separate lake basins indicates some cause-and-effect relationship, such as a connection between the basins."

Hough then proceeds to find such a connection across the northern part of the Southern Peninsula of Michigan along the retreating margin of the Valders ice in the Lake Michigan Basin. A map of this postulated water connection is shown in figure 38 on page 180. Such a correlation requires the persistence of the Glenwood stage for a much longer time than previously determined by Bretz who considered the Glenwood stage to be no younger than the Port Huron moraine. Hough's interpretation also demands northward rather than southward flowing water through the Port Huron outlet during both the Grassmere and Lundy stages.

The author presents a good deal of evidence in behalf of his thesis that the waters of Lake Algonquin were restricted to the Lake Michigan and Lake Huron Basins and that the shorelines in the Lake Superior Basin formerly attributed to Lake Algonquin by Leverett were in fact related in part to Lake Duluth instead. During Algonquin time, glacial ice still occupied the eastern end of the Lake Superior Basin according to Hough's interpretation (p. 219).

One section of the book which will be valuable to Pleistocene geologists in the Great Lakes region is the chapter on radiocarbon chronology of Great Lakes history (p. 269). In this chapter the author has gathered together all radiocarbon dates from the literature into tables for useful reference. This catalogue gives a full description of each sample, its laboratory number, age in C-14 years, geological assignment by the author, and reference where first published. It is refreshing to note that Hough uses radiocarbon dates to provide absolute ages for the various lake stages *after* the geological evidence

has been considered rather than trying to build a sequence of events on radiocarbon dates alone.

The lake stage maps (figs. 53-75, p. 284-296) provide a thumbnail sketch of the interpretation of lake history envisioned by Hough and will undoubtedly stand for years to come as the standard replacement of the old color plates of Leverett and Taylor.

The bibliography cites 234 references, ten of which are papers by the author himself. The index contains only 192 entries while the table of contents lists 157. The book is printed on a good grade of paper in fairly large type.

Geologists, ecologists, archeologists, and even thte informed lay reader will find this book an important contribution to the knowledge of the Great Lakes. The author is to be congratulated for his forthrightness of presentation and his courage in expressing new ideas that are not in accord with time-honored concepts. Hough may be accused of sticking his neck out in some places in the book, but he can never be accused of straddling the fence on a controversial subject. The reader is never in doubt about Hough's position, and whether one agrees with his interpretation or not, one cannot help but admire the freshness of many of his ideas. Certainly, *Geology of the Great Lakes* is a must for the bookshelves of all who are concerned with these ever intriguing water bodies which constitute the largest fresh-water system in the world.

JAMES H. ZUMBERGE

ERRATUM

The figure that appears above the caption "Fig. 5" on page 179 of the March issue (v. 257, no. 3) is figure 9 of the article by Ph. H. Kuenen (p. 172-190). Similarly, the figure that appears above the caption "Fig. 9" on page 187 is figure 5 of the same article.

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